PAPERS IN PHYSICAL OCEANOGRAPHY AND METEOROLOGY

PUBLISHED BY

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

AND

WOODS HOLE OCEANOGRAPHIC INSTITUTION

(In continuation of Massachusetts Institute of Technology Meteorological Papers)

VOL. III, NO. 4

THE CYCLE OF PHOSPHORUS IN THE WESTERN BASIN OF THE NORTH ATLANTIC

I

Phosphate Phosphorus

BY

H. R. SEIWELL

Contribution No. 64 from the Woods Hole Oceanographic Institution

CAMBRIDGE, MASSACHUSETTS

April, 1935

CONTENTS

INTRODUCTION

The importance of phosphorus for organic production in the sea appears to have been recognized first by Brandt (1899) and the earlier determinations of this element in the coastal seas of northern Europe (Brandt, 1920; Raben, 1920; Mathews, 1917) suggested a correlation between seasonal variation of phosphate and growth of phytoplankton. These earlier determinations were later shown to be too high (Atkins, 1926, a) and did not indicate the complete exhaustion of phosphate from the water, so it was not until several years later that Atkins (1923), employing the rapid and more accurate colorimetric ceruleo-molybdate method of Denigès, illustrated the complete dependence of algal growth on phosphate (in the English Channel) and thus established the foundation for modern studies of marine chemical fertility.[1]

The beginning of our knowledge of phosphate content of the open ocean may, as far as is known to me, also be attributed to Atkins (1926, a) and even though these early results were frequently somewhat vitiated by storing of the samples before analyses, they represented the order of magnitude of phosphate concentration in the sea. Within recent years phosphate determination has become a component part of the program of most deep sea investigations and much general information on its distribution and variation in the open ocean has been brought to light.

From a geographical standpoint the most widespread investigation of phosphate in the oceans (Atlantic, Pacific, and Indian) was made on the recent cruise of the "Dana" (Thomsen, 1931); and this together with the researches from the various "Discovery" cruises, principally in the South Atlantic (Deacon, 1933); by Ruud (1930) on the whale factory s/s "Vikingen" in the Weddell Sea; by the "Meteor" both in the South Atlantic (Wattenberg, 1933) and northern North Atlantic (Böhnecke, et al, 1932); by the "Carnegie" in the North Atlantic and Pacific (Moberg, et al, 1930; Seiwell, 1931); by the "Nautilus" in the high latitudes of the North Atlantic (Sverdrup, 1933); and by the "Atlantis" in the western North Atlantic have all contributed a wealth of material to our knowledge of phosphate distribution in the ocean basins.[2,3]

The material for this report has been obtained principally from "Atlantis" cruises in the western North Atlantic and it is planned to follow the present account with a second part treating other phases involved in the phosphorus cycle of this region.

Professor Henry B. Bigelow has guided the preparation of this report and Professor C.-G. Rossby has been consulted on the theoretical discussion.

[1] See Atkins (1926, a) for references to earlier work.

[2] References to these cruises are not complete and merely serve as a guide to more detailed information.

[3] A number of important studies of the phosphate content of coastal waters of northern Europe and North America have been made during the past decade. For bibliography and references see especially: Gran and Thompson (1930); Gran (1932); Braarud and Klem (1931); and Buch (1932).

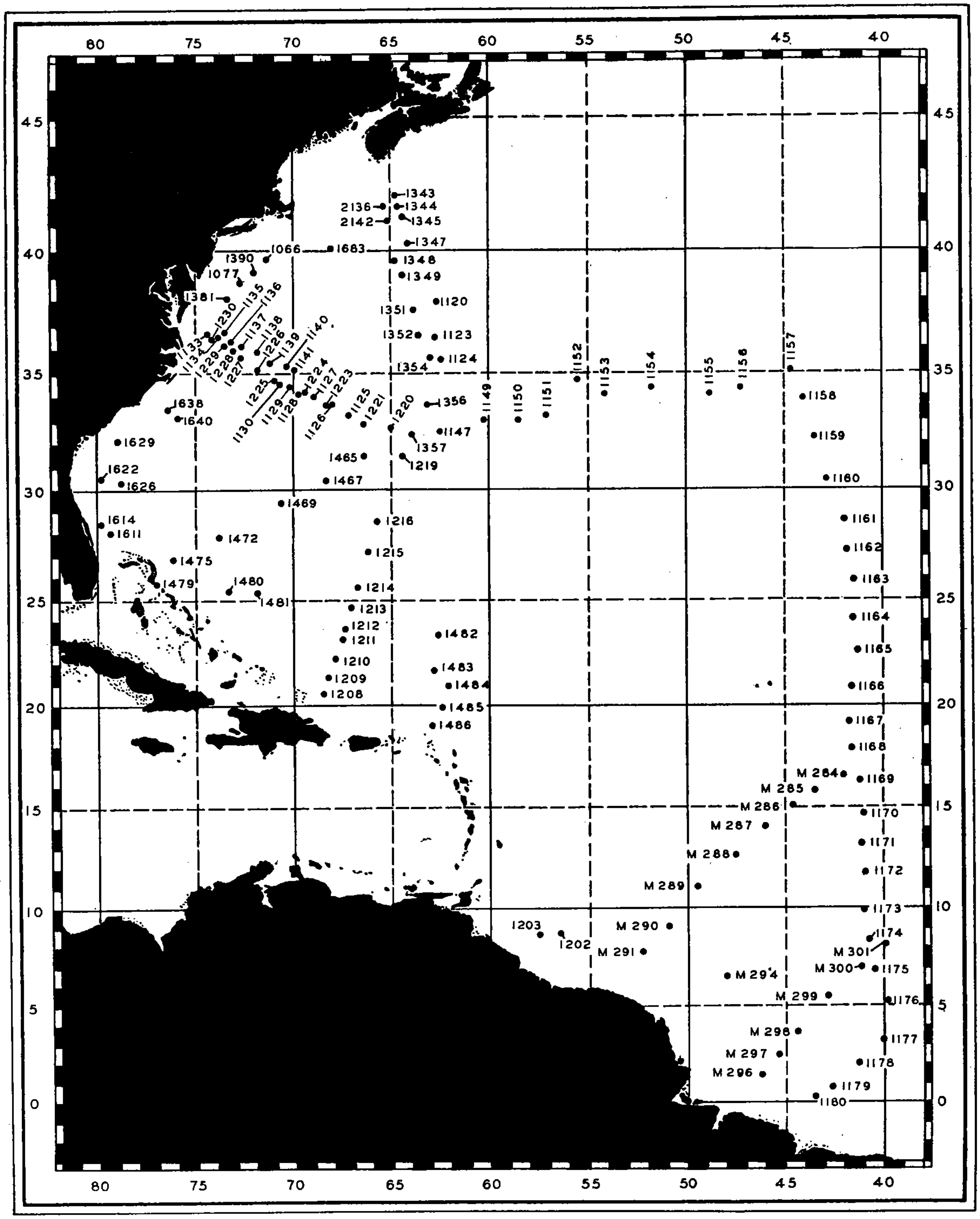

FIG. 1. Chart of area of investigation in western North Atlantic. All stations were made by "Atlantis" except M stations by "Meteor" (Wattenberg, 1933).

REGION OF INVESTIGATION

This discussion is confined to the western basin of the North Atlantic (fig. 1) and is based principally on observations obtained during the following cruises of "Atlantis":

Bermuda eastward to 35th meridian: stations 1147–1157, February 24 to March 4, 1932.

Latitude 35°N to equator, along 40th meridian: stations 1157–1179, March 4–21, 1932.

Haiti to Bermuda: stations 1208–1219, April 7–13, 1932.

Bermuda to Chesapeake Bay: stations 1134–1142, February 12–15, 1932; and stations 1220–1231, April 17–23, 1932.

Nova Scotia to Bermuda: stations 1343–1357, August 14–21, 1932.

Miscellaneous "Atlantis" stations: 1465–1486 and 1611–1640 made during the late winter and spring of 1933.

The "Atlantis" observations are published in the Bulletin Hydrographique (1933) without correction for salt or temperature error and the corrected data on which this report is based (page 8) is graphically presented herein and can be scaled with a fair degree of accuracy.

METHODS

With slight variations adapted to fit the particular conditions,[4,5] the "Atlantis" phosphate determinations for 1931 and 1932 were made by the colorimetric method of Denigès as modified by Atkins (1924).

Reagents

1. Ammonium molybdate-sulphuric acid mixture. A 10 per cent solution of ammonium molybdate is diluted with three times its volume of 50 per cent (by volume) sulphuric acid and kept in dark glass bottles; 2 cc are required for 100 cc of sample.

2. Stannous chloride solution. This solution is prepared fresh for each set of analyses by dissolving about 0.1 gram of tin foil in 2 cc of concentrated hydrochloric acid with one drop of 3 to 4 per cent copper sulphate solution and then diluting to 10 cc. Depending on the phosphate concentration of the sample, 1 to 3 drops are required per 100 cc of sample.

3. Standard phosphate solutions. A series of standard phosphate solutions with concentrations ranging from 6.2 to 62.0 milligrams of P per cubic meter, in 6.2 milligram intervals, were prepared for each series of analyses by diluting appropriate amounts of M/50,000 KH_2PO_4 with distilled water to 100 cc; the standards and samples are treated simultaneously with reagents.

[4] The 1933 "Atlantis" phosphate observations in the western basin of the North Atlantic (comprising a very small part of the total number) were made by investigators who used procedures differing from that described here (see Bulletin Hydrographique pour L'Annee, 1933).

[5] Also see Harvey (1928 and 1929) for discussion of method of phosphate estimation in sea water.

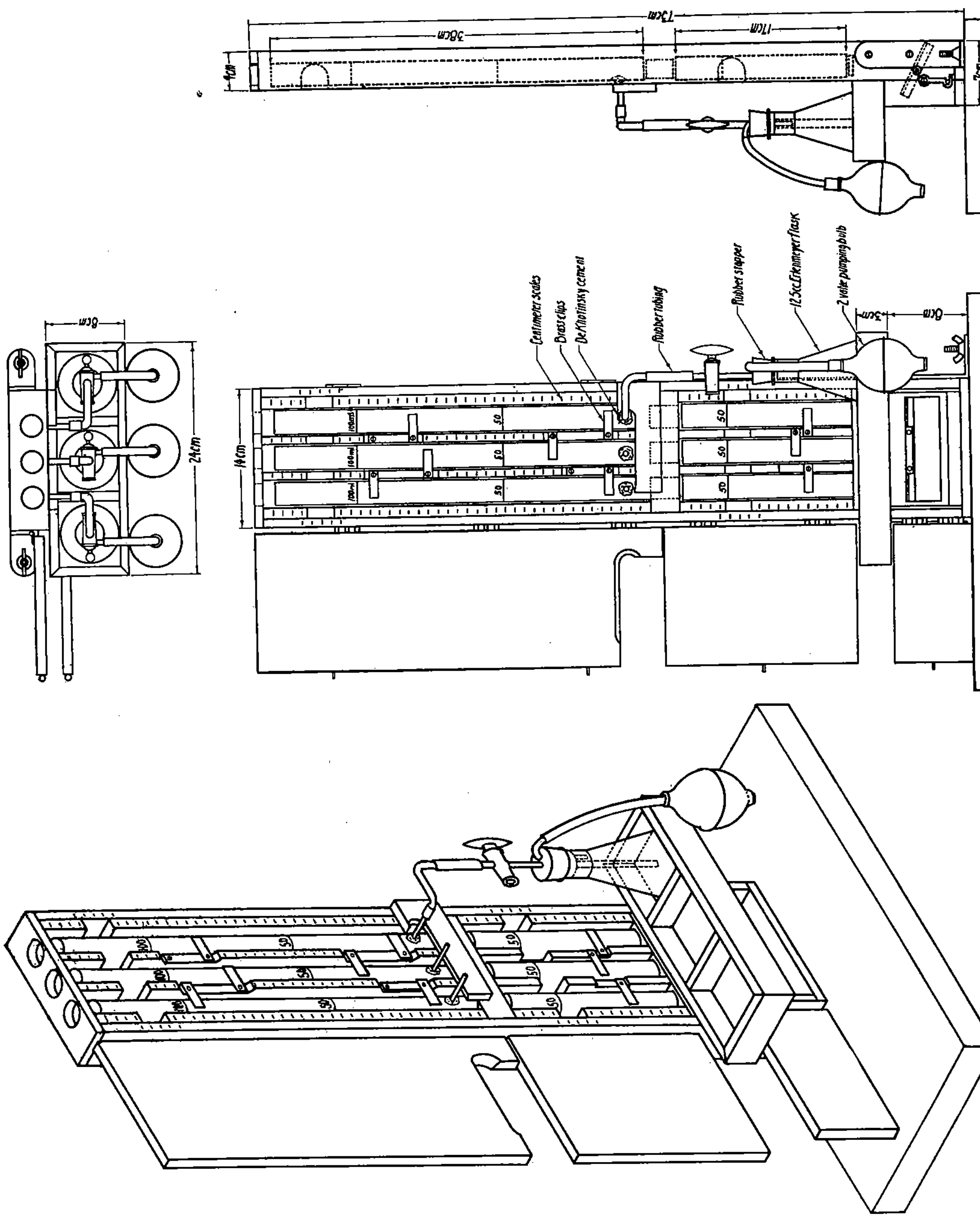

FIG. 2. Design of apparatus for colorimetric estimation of phosphate. See text, page 7.

Colorimeter

Figure 2 shows the design and construction of the colorimeter used on "Atlantis" from October 1932 to September 1933. The instrument and a colorimeter lamp were bolted to a firm table in a semidarkened corner of the laboratory. Color comparisons are made in the upper set of three uniform color comparison tubes (sample in center with consecutive standards on either side); this arrangement allows several readings to be made rapidly on each sample. The lower tubes are filled with clear distilled water unless the sea water sample (before adding reagents) is turbid or discolored in which case the color balance may be restored by placing a sea water sample without reagents in each tube below those containing the standard phosphate solutions.

Procedure

Phosphate analyses were completed within a few hours after collecting the samples and frequently it was necessary (because of other duties) to make determinations before the sea water samples had warmed up to laboratory temperature. During 1931 and 1932 temperatures of both standards and samples at time of comparison were recorded so that whenever necessary the results have been corrected according to the following equation:

$$P = P' + k_t(t_s - t_w)P'$$

where P is corrected phosphate value; P', the estimated phosphate value according to Beer's law; t_s, temperature of phosphate standard; t_w, temperature of sample at time of comparison; k_t, a coefficient for temperature influence on intensity of the phosphomolybdate blue (0.012 to 0.016 for conditions of this investigation, Brujewicz and Krasnova, 1933). By applying this temperature correction the original phosphate result may be increased 10 to 20 per cent and in a few extreme cases even higher; in general, greater corrections are applied to results from the great depths.

The data have not been corrected for salt error except as noted for certain purposes of calculation. The effect of sea salt on the colorimetric estimation of phosphate in sea water has been studied by Kalle (1934), Ibañez (1933), and Brujewicz and Krasnova (1933); whenever required the adjustment factor (1.34 to 1.36 for salinities between 32 o/oo and 38 o/oo) proposed by Brujewicz and Krasnova has been used.

Accuracy of Results

Because of the variable factors which may cause errors in phosphate analyses at sea it is not possible to assign definite limits of accuracy to the results on which this discussion is based. However, it is reasonable to assume (since most of the observations contained herein were made by a single investigator using the same apparatus and the same method) that the results, in general, have a fair relative correctness. And as they are in agreement with the few observations made by other observers in the area of investigation (page 33) it seems that the order of magnitude of correctness is about the same as that for customary analyses of this kind when made at sea. The question of variability of the results is discussed later (pages 21 and 27).

The ability of the individual observer to obtain reasonably consistent results from colorimetric analyses at sea (assuming that the procedure and apparatus is correct) depends largely on the variable sensitiveness of vision to different concentrations of color and on the amount of error which may arise from fatigue and eyestrain. Eyestrain in particular may become quite prominent as frequently the investigator must work long continuous periods to complete the analyses and the desire to obtain a large number of observations may sometimes unknowingly result in a sacrifice of accuracy. On the other hand, among different observers working under dissimilar conditions divergence of analytical results are liable to result both from the above factors as well as from the use of different technic and equipment, and from the varying degrees of correctness by which different individuals judge colors.[6] However, by the use of carefully worked out analytical procedures, and by the calibration of colorimeters, with mechanical and optical defects kept at a minimum, much can be done to eliminate errors.

[6] For discussion of fallacies in colorimetry see Dehn (1917).

VERTICAL DISTRIBUTION OF PHOSPHATE PHOSPHORUS

In vertical section the phosphate content of the western basin of the North Atlantic is lowest (usually 0–4 mg P per m^3) in the more or less homogeneous water overlying the thermocline[7] (<400 to >40 meters thick) below which it increases to a maximum value (<30 to >60 mg P m^3) at intermediate depths of usually 800 to 1000 meters. In still deeper water phosphate content either decreases or remains relatively constant so that at 2000 meters the usual horizontal range is 25 to 35 mg P per m^3. Significant phosphate gradients below depths of maximum concentration occur only when the maximum is high, e.g., in the southern half of the area. Within the area of investigation variations in the vertical phosphate gradient are principally due to: (1) variation in thickness of the phosphate poor surface layer; and (2) variation in value and depth of the maximum phosphate concentration. In order to bring out regional differences in the vertical distribution of phosphate, conditions as existing along the several "Atlantis" sections at the times of observation are considered separately.[8]

Northeastern Sargasso Sea Between Bermuda and 35th Meridian[9]

The distribution of phosphate along this section (fig. 3) is typical of the region which it crosses (fig. 1). During February 24 to March 4, 1932 the essential features to be stressed were as follows. The surface phosphate content was 1 to 3 mg P per cubic meter (table 1) and the phosphate poor layer, containing less than 5 mg P per cubic meter, extended to depths of 350 to 550 meters, below which phosphate increased to its maximum value (about 30 mg P per m^3; table 1) at intermediate depths of 800 to 1000 meters. In the still deeper water phosphate content declined but little with increasing depth (at 2000 meters values were usually between 25 and 27 mg P per m^3) so that the maximum values at intermediate depths are not well defined, but merely mark out the lower limit of the increasing phosphate gradient in the upper part of the water column (fig. 4).

At the time of observation the surface temperature along this section was probably near its annual minimum (17.69°–19.32°; table 1) and the water overlying the principle thermocline was nearly homogeneous (fig. 5). The depths at which the thermocline began are somewhat difficult to determine as there is no sharp distinction between it and the overlying water strata, but the first significant change of curvature on the temperature-depth graphs (fig. 5) usually began between 300 and 400 meters depth. As the thick overlying homogeneous layer of water in this region is probably an accumulation

[7] In this paper the depths designated as the beginning of the thermocline are those at which permanent significant decreases of temperature appear to begin, which are assumed to represent the maximum depths reached by homogeneous water.

[8] Phosphate values in this discussion are not corrected for salt error unless specifically indicated (page 7).

[9] Observed salinity range for this section is 34.91 o/oo to 36.69 o/oo; factor for correcting phosphate values for salt error is 1.35, according to Brujewicz and Krasnova (1933).

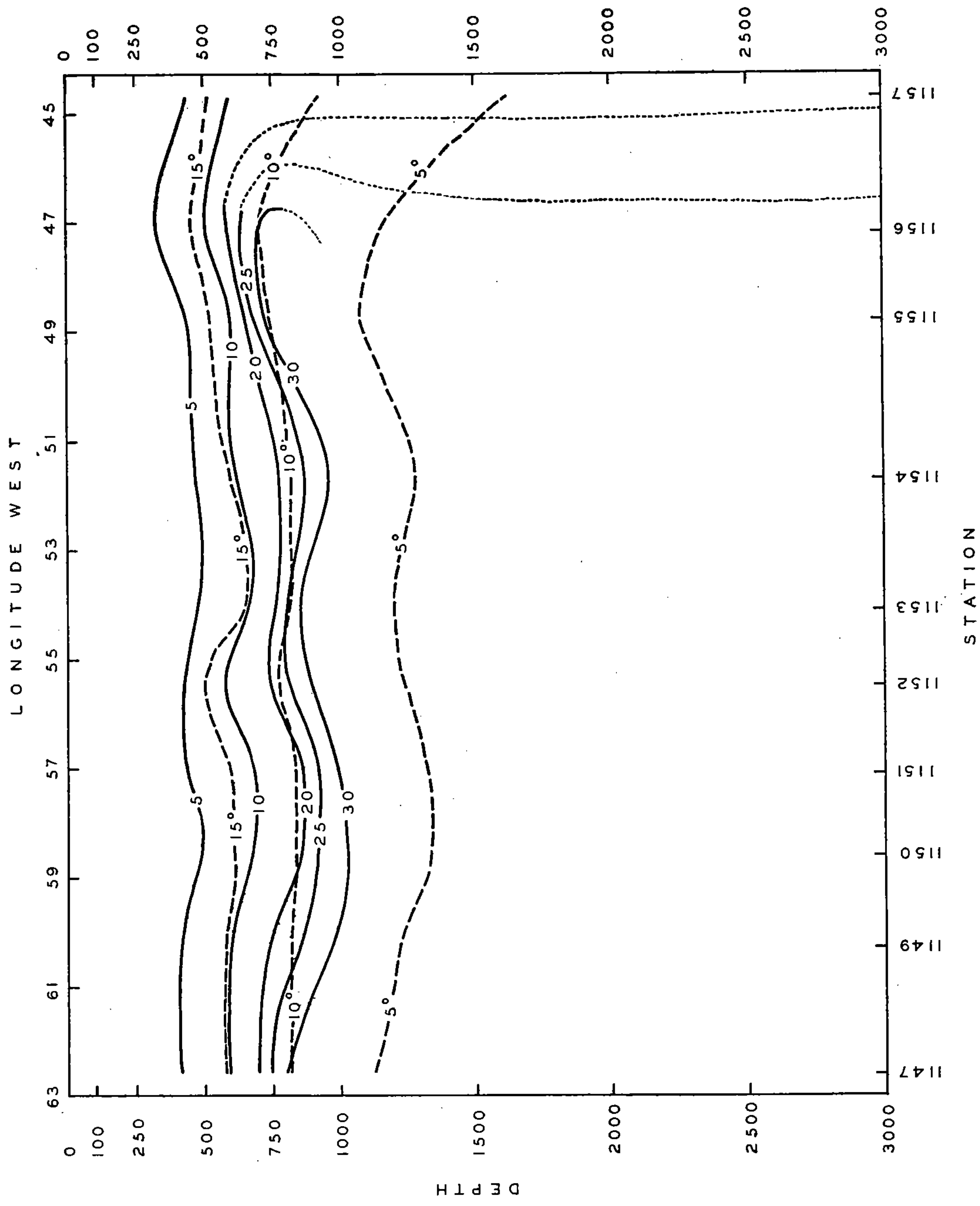

FIG. 3. Distribution of phosphate, milligrams of P per cubic meter, "Atlantis" stations 1147–1157, longitude 62°35′W to 44°40′W between latitude 32°37′N and 35°10′N; February–March 1932.

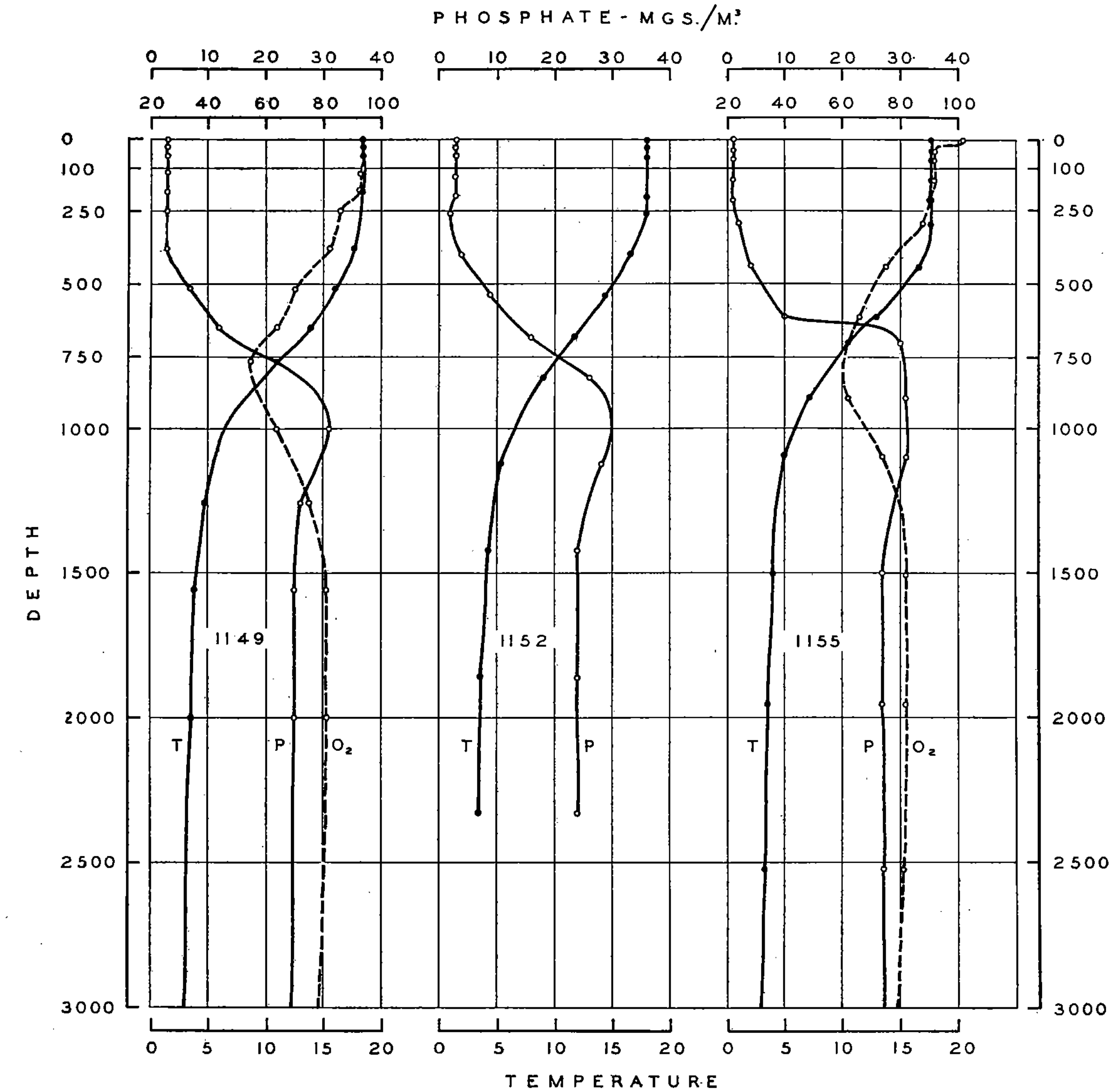

FIG. 4. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in northern Sargasso Sea; February–March 1932. For station positions see table 7.

of surface water it is understandable that its phosphate content should be very low (1–4 mg P per m³) and its oxygen content very high (figs. 4, 5). The small but distinct oxygen and phosphate gradients sometimes occurring in the lower part of the homogeneous layer (see stations 1149, 1150, and 1151; fig. 5) are presumably due to a concentration of products resulting from decomposition of organic materials below the depth of plant activity which have not yet been redistributed throughout the entire layer. That complete stirring of the homogeneous layer occurs in winter appears to be indicated by the uniform vertical gradients of phosphate, oxygen and temperature as at stations 1152, 1153, and 1154 (fig. 5). Mixing of the homogeneous layer does not appear to extend into the phosphate rich layers of the thermocline (fig. 4) but will bring within range of plant organisms nutrient substances which may by various means tend to accumulate above the thermocline. Resistance of the water overlying the main thermocline to vertical mixing is greatly increased during summer by the development of an upper temporary thermocline about 100 to 150 meters thick due to warming of the surface layers, illustrated by station 1041 (36°55′N, 52° 41′W; August 15, 1931; fig. 5).[10]

Throughout this section significant increases of phosphate and decreases of oxygen content begin at about the same depths as the thermocline; phosphate remains relatively constant below the depth of its maximum concentration whereas oxygen content, after reaching its minimum, reverses its gradient and increases in the still deeper water so that the water below 2000 meters contains large amounts both of phosphate and of oxygen (table 1; fig. 4).

[10] In August (1931) at station 1041 (36°55′N, 52°41′W) the average vertical variation of σ_t in the 0–100 meter layer was 2.1×10^{-2} units per meter; whereas during the following February (1932) at station 1154 (34°20′N, 51°45′W) the σ_t variation in this layer was zero.

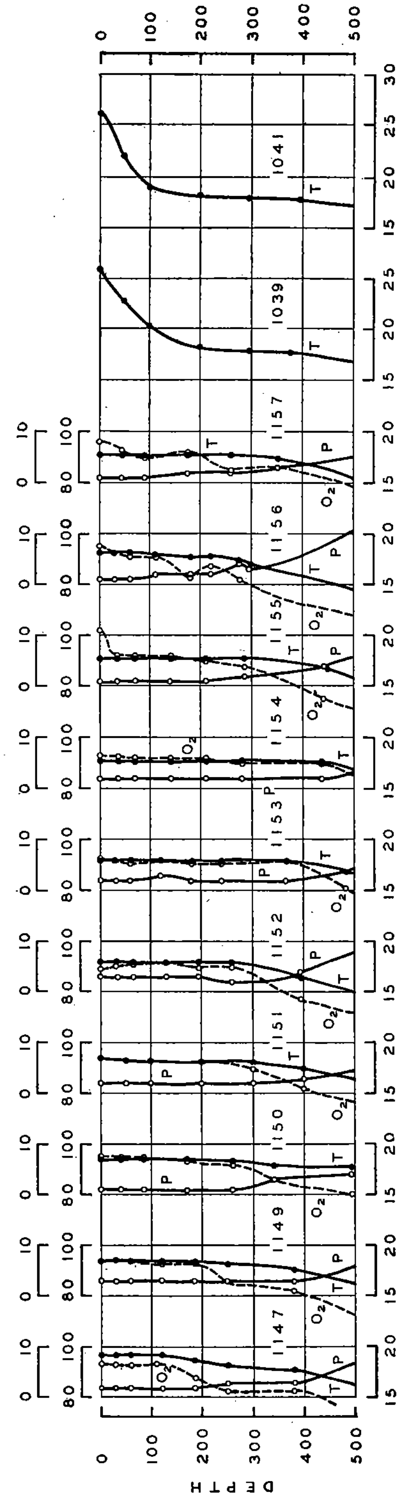

FIG. 5. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in the upper 500 meters in northern Sargasso Sea; February–March 1932. For station positions see table 7.

TABLE 1

Station	Surface		Thermocline Beginning			P Gradient End		2000 Meter
	T°	P Mg/M³	Depth	T°	P Mg/M³	Depth	P Mg/M³	P Mg/M³
1147	19.32	2	382	17.76	3	808	33	27
1149	18.42	3	384	17.63	3	1002	31	25
1150	18.49	1	496	17.82	4	937	27	(25)
1151	18.48	2	400	17.50	3	1000	31	(27)
1152	17.89	3	262	17.88	2	821	28	24
1153	17.94	2	363	17.90	2	836	30	25
1154	17.76	2	435	17.65	2	934	27	(26)
1155	17.69	1	285	17.71	2	892	31	27
1156	18.15	1	296	16.94	3	758	30	(27)
1157	17.80	1	351	17.38	3	1003	18	18

"Atlantis" stations between Bermuda and 45th meridian, February–March 1932. Thermocline beginning estimated from observed data (see page 9); 2000 meter values scaled from station curves; bracketed values estimated. End of phosphate gradient means depth at which rapid increases of phosphate content with increasing depth cease as determined by direct observation. This is also depth and value of phosphate maxima, except at station 1151 where a concentration of 32 mg P per cubic meter was observed at 1200 meters depth. For station positions see table 7.

Mid Atlantic along 40th Meridian between Latitude 35°N and Equator[11]

The distribution of phosphate along this section changes from north to south (fig. 6); at the time of observation (March 4–21, 1932) the essential features were as follows. The surface concentration ranged from 0 to 4 mg P per cubic meter; and the phosphate poor layer, containing 5 mg P or less per meter, extended to about 400 meters depth north of latitude 26°N, to the south of which it gradually diminished in thickness to 25 meters near latitude 8°N; increasing again to about 100 meters near the equator. In deeper water phosphate content increased rapidly with increasing depth to a maximum value at intermediate depths; north of 27°N this maximum was 30 mg P or less per cubic meter at about 900–1000 meters; south of 27°N it increased up to 60 mg P per cubic meter and, in general, the rapidly increasing phosphate gradient ended 100 to 300 meters closer to the surface than it did north of 27°N. In the still deeper water phosphate tends to approach a uniformity. At 2000 meters depth it was usually in the neighborhood of 30 mg P per cubic meter; so that with increased concentration at intermediate depths the phosphate maximum is better defined in the south (table 2; figs. 6, 7).

Table 2, figures 6 and 7 illustrate that in the northern part of the section (stations 1157, 35°10′N, to 1160, 30°26′N) the phosphate content below the phosphate poor layer is much lower than the average for the remainder of the area of investigation. At this time it is not possible to judge if the difference is real or represents an observational error. But it is noteworthy that the phenomenon occurs in that part of the area where Iselin[12] found a well defined salinity anomaly (presumably indicating considerable admixture of Mediterranean water); and since Thomsen (1931) has shown that Mediterranean water is very poor in phosphate, ranging from 2 to 15 mg P per cubic meter (5 to 35 mg P_2O_5 per cubic meter) between 1000 and 4000 meters depth an explanation of the phenomenon may be inferred.

At the time of observation surface temperature of this section was, no doubt, close

[11] Salinity range observed in this section is 34.51 o/oo to 37.34 o/oo; factor for correcting phosphate values for salt error is 1.35–1.36 (Brujewicz and Krasnova, 1933).

[12] Personal communication.

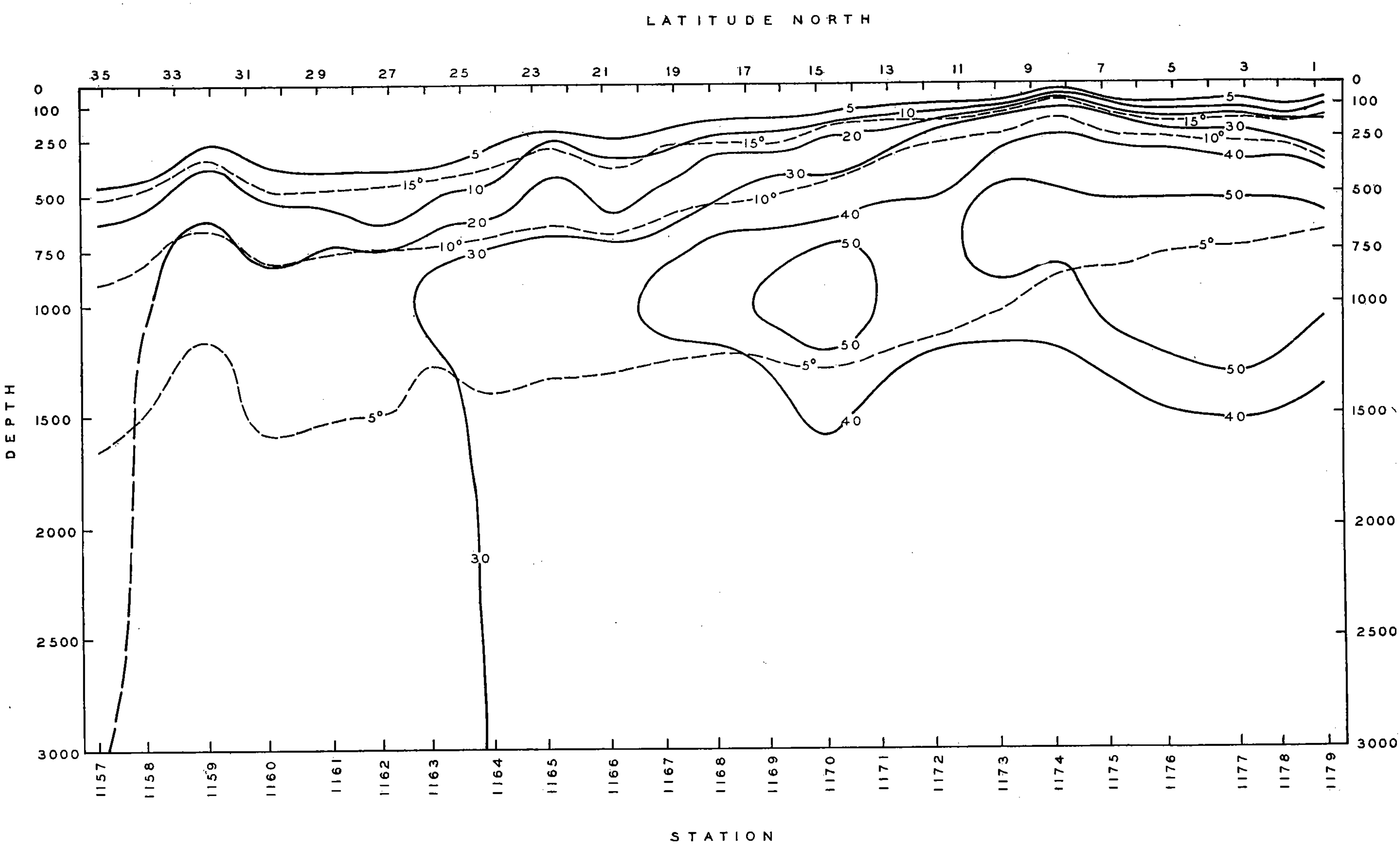

FIG. 6. Distribution of phosphate (mg P per m³) in mid Atlantic along 40th meridian between latitude 35°N and equator; March 1932. For station positions see table 7.

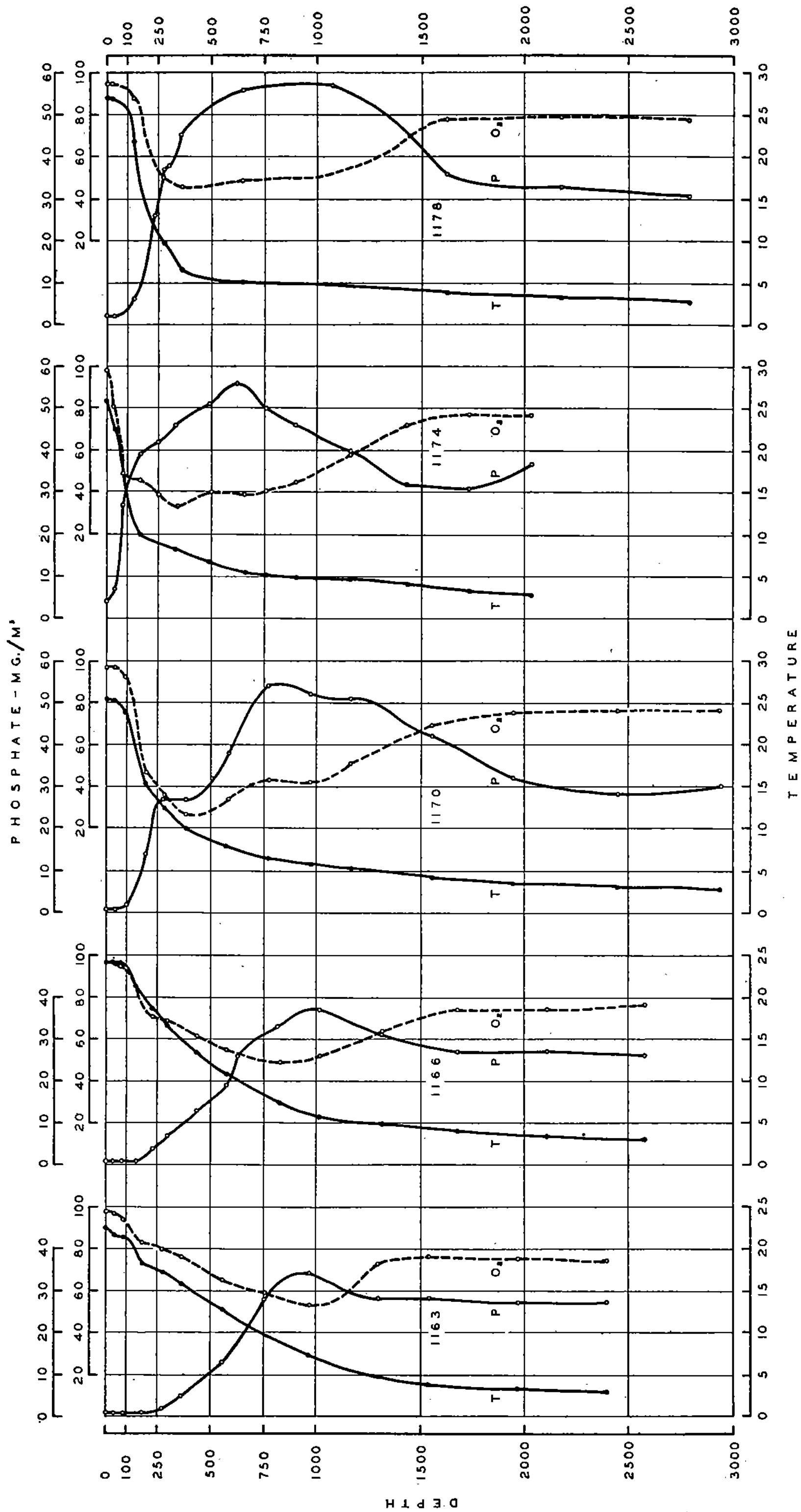

FIG. 7. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in mid Atlantic along 40th meridian; March 1932. For station positions see table 7.

to its annual minimum so that the water overlying the thermocline showed its maximum homegeneity. Fig. 8 illustrates that there is considerable variation of depth at which the thermocline begins along the section: the depths of thermocline given in table 2 are those at which the first significant changes of curvature occurred on the temperature-depth graphs. Thus, north of latitude 32°N (sta. 1159) the thermocline begins more than 200 meters below the surface and is not well marked off from the overlying water, while south of latitude 27°N it is always less than 100 meters from the surface and loosely demarked at first but becoming more distinct toward the south. Comparison of the vertical distribution of oxygen, phosphate, and temperature in the upper part of the water column (fig. 8) shows that north of latitude 32°N (sta. 1159) significant changes in vertical distribution of all three elements begin at about the same depths (below the thick homogeneous layer); further south, in the north central part of the section (stas. 1160 to 1166, latitude 30°26′N to 20°50′N), where the temperature discontinuity layer begins within 100 meters of the surface but is not sharply marked out from the overlying water, the significant increases of phosphate content begin deeper than do decreases of temperature and oxygen; however, still further south (south of station 1166), as the upper boundary of the thermocline becomes more distinct, definite increases in phosphate content parallel decreases in oxygen and temperature.

TABLE 2

Station	Surface		Thermocline Beginning			P Gradient End		2000 Meter
	T°	P Mg/M³	Depth	T°	P Mg/M³	Depth	P Mg/M³	P Mg/M³
1157	17.80	1	351	17.38	3	1003	18	18
1158	17.74	2	281	17.44	2	752	19	21
1159	18.04	2	208	17.06	4	732	26	21
1160	20.38	1	154	19.88	2	954	23	20
1161	20.81	1	74	20.80	2	931	25	25
1162	22.04	2	100	21.80	1	995	27	25
1163	22.47	1	92	21.37	1	967	34	27
1164	22.90	1	95	22.19	1	990	39	31
1165	23.41	2	99	23.00	1	795	33	29
1166	24.26	1	75	24.10	1	1024	37	27
1167	24.50	1	93	24.00	1	882	45	32
1168	24.69	1	94	24.19	1	759	48	29
1169	25.07	2	100	23.09	2	993	54	29
1170	25.45	1	97	23.86	2	784	54	32
1171	25.51	0	80	25.19	0	778	49	30
1172	25.60	0	76	25.55	0	662	48	29
1173	25.64	2	40	25.70	1	629	54	29
1174	25.73	4	<40	>22.62	(5)	634	56	37
1175	27.08	3	84	25.01	2	936	59	29
1176	26.91	3	85	26.90	3	852	59	(33)
1177	27.08	4	53	26.90	2	817	61	(32)
1178	27.05	2	—	—	—	664	56	31
1179	27.50	3	47	27.25	2	890	54	34

"Atlantis" stations along 40th meridian between latitude 35°N and equator. Thermocline beginning estimated from observed values (see page 16); 2000 meter values scaled from station curves; bracketed values estimated. At station 1178 observations in upper part of water column were not sufficiently close to estimate beginning of thermocline. End of phosphate gradient is estimated depth at which rapid increases of phosphate content, with increasing depth, cease as shown by observation. This is also depth and value of phosphate maxima. For station positions see table 7.

The relationship of phosphate and oxygen content of the water overlying the thermocline to the depth of plant activity is discussed in a later chapter (page 50). The fact that oxygen decreases conformably with temperature seems to indicate that most plant activity occurs above the temperature discontinuity layer, even when the latter begins

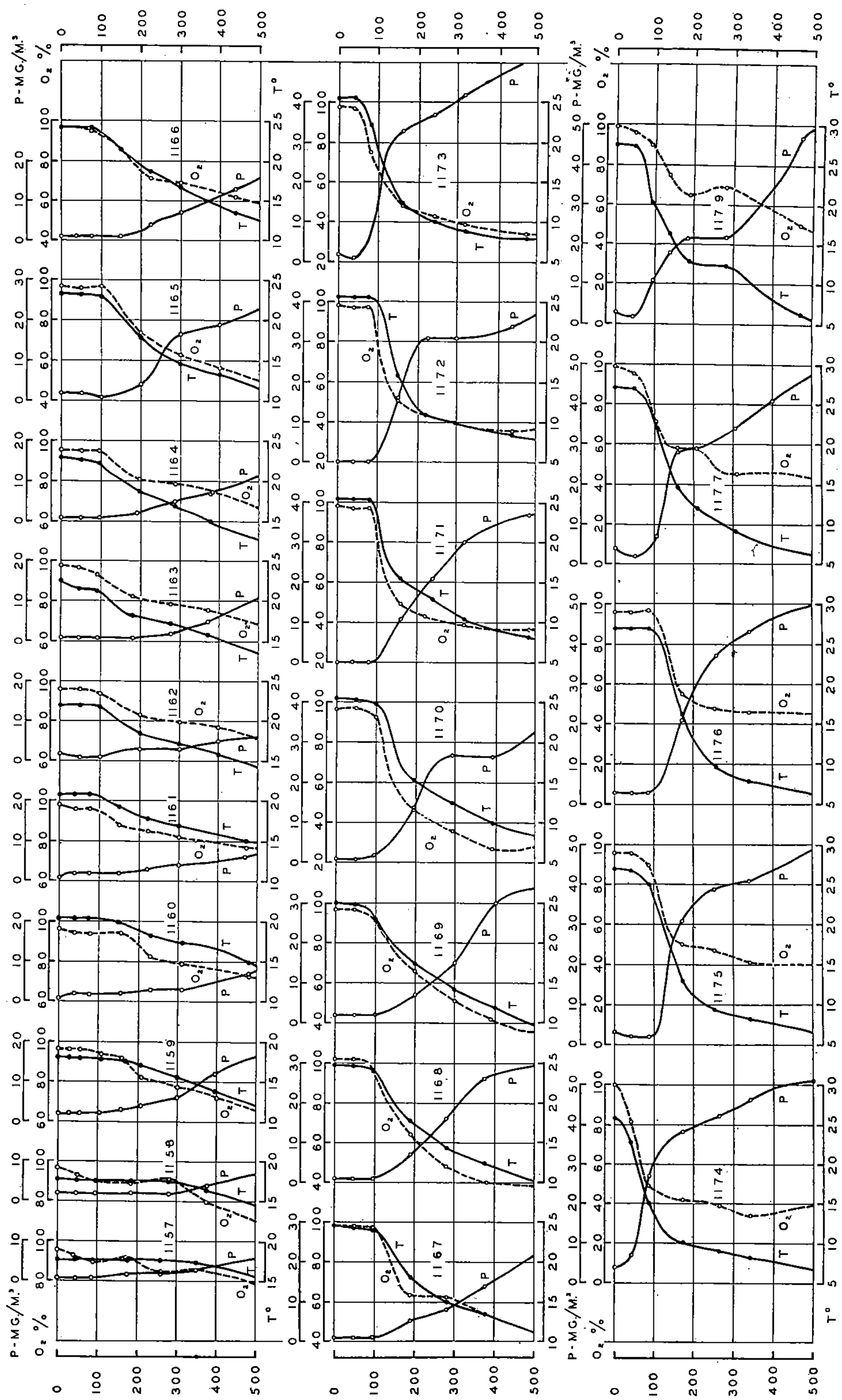

FIG. 8. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in the upper 500 meters of mid Atlantic along 40th meridian between latitude 35°N and equator; March 1932. For station positions see table 7.

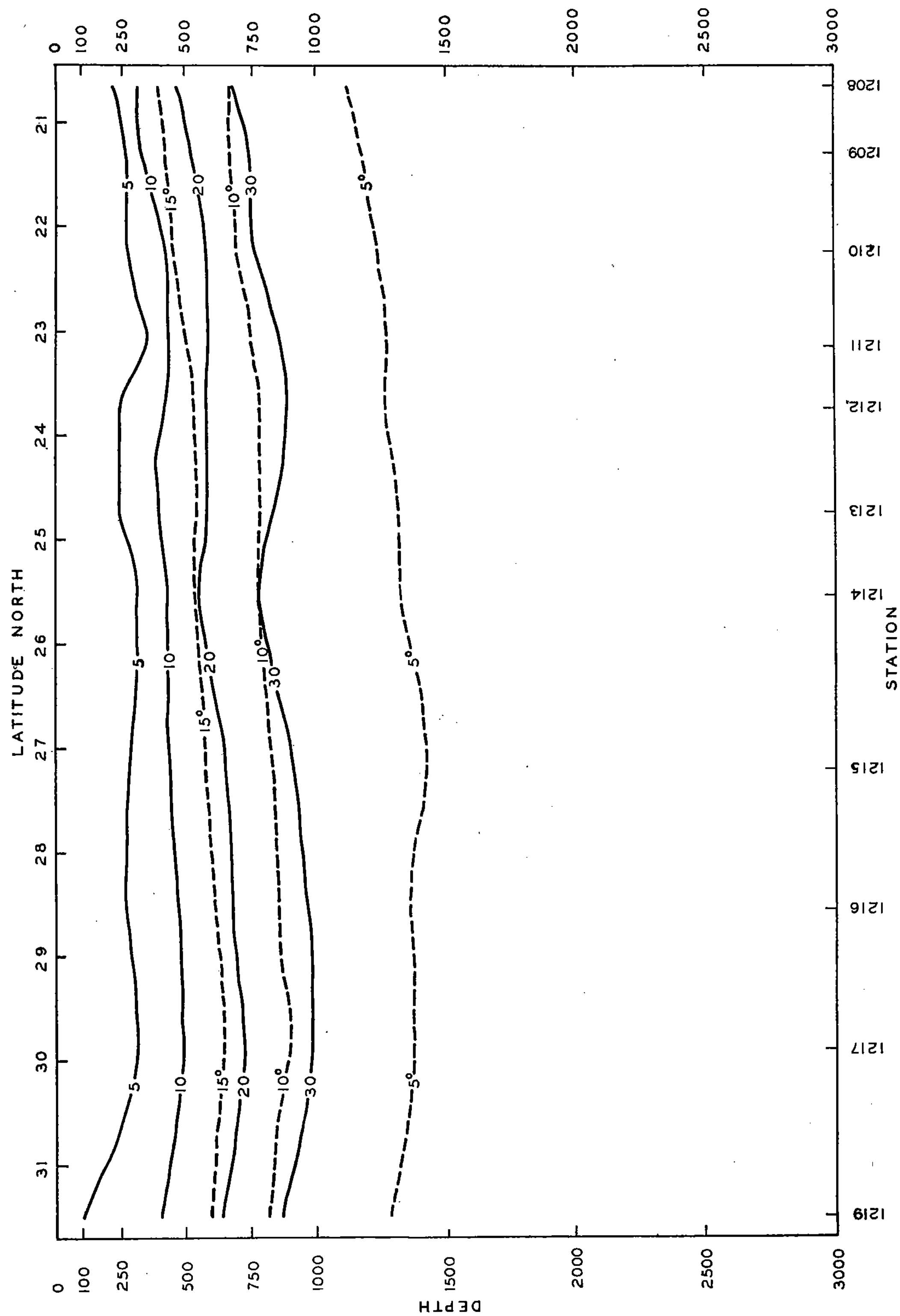

FIG. 9. Distribution of phosphate (mg **P** per m³) between Haiti and Bermuda; April 1932. For station positions see table 7.

less than 40 meters from the surface. In the north central part of the section (stas. 1160 to 1166) the extension of the phosphate poor layer into the thermocline does not appear

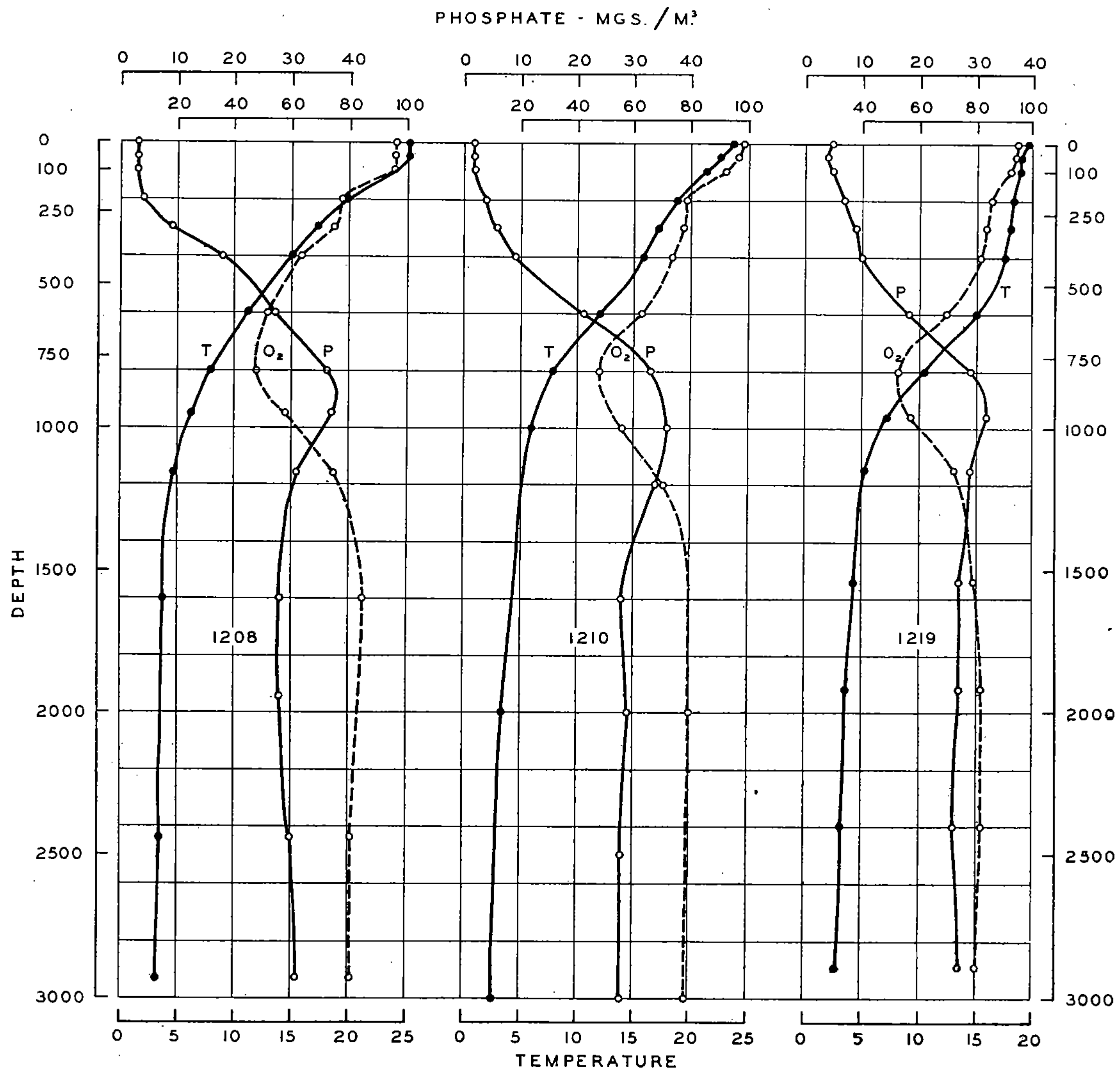

FIG. 10. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature between Haiti and Bermuda; April 1932. For station positions see table 7.

to be due to phytoplanktonic activity in situ since the vertical distribution of relative oxygen content, which decreases significantly along with the temperature, suggests otherwise (see sta. 1163; fig. 8).

Between Haiti and Bermuda[13]

The distribution of phosphate along this section as determined in the early spring of 1932 (April 7 to 13) illustrated by figure 9, shows surface phosphate values (usually 1 to 3 mg P per m³) and depth of phosphate poor layer (250–300 meters) to be about the same as described for similar latitudes (32° to 20°N) further east (fig. 6; page 13). The increasing phosphate gradient ended at depths of 800 to 1000 meters, the higher maximum values characterizing the southern part of the section; and in still deeper water phosphate content declined somewhat with increased depth until at 2000 meters it is between 25 and 29 mg P per cubic meter (table 3; figs. 9, 10).

At the time of observation definite warming of the surface layers appears to have set in so that the upper limit of the main thermocline had become indistinct. However, as the high and relatively uniform oxygen and low phosphate contents which characterize the water overlying the main thermocline persisted (fig. 11) we were enabled to estimate the beginning of the main thermocline in doubtful cases, the depths of which are tabulated in table 3. These estimations may be in error 50 meters or more, but conditions in the upper part of the water column (fig. 11) are similar to those at the same latitudes further east (fig. 8) in that the upper limit of the main thermocline is not distinct from the overlying water and in that the phosphate poor water extends well into the thermocline. Depth and concentrations of the maximal phosphate content of the midstrata are also similar.

[13] Observed salinity range in this section is 34.90 o/oo to 36.82 o/oo; factor for correcting phosphate values for salt error is 1.35–1.36 (Brujewicz and Krasnova, 1933).

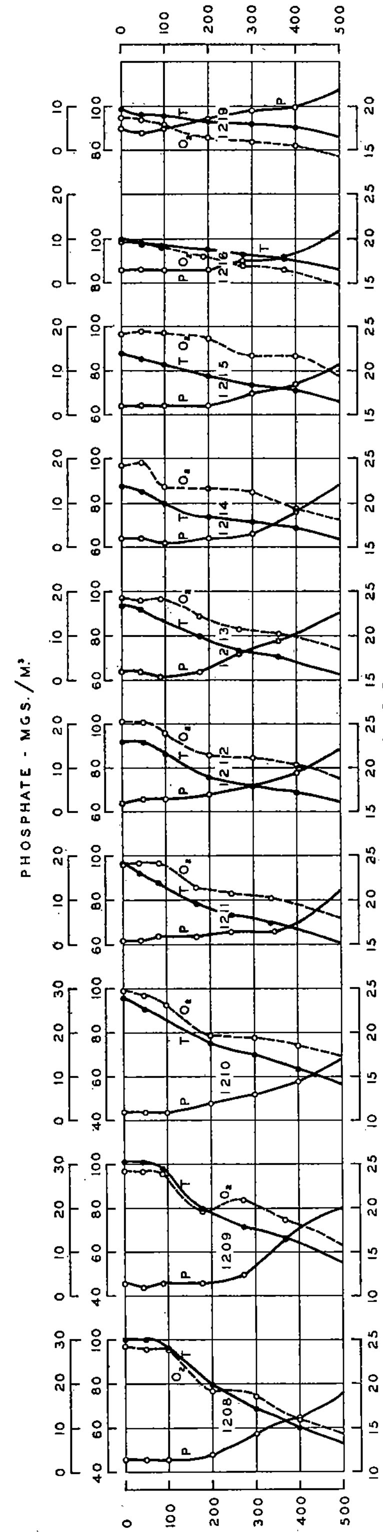

FIG. 11. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in the upper 500 meters between Haiti and Bermuda; April, 1932. For station positions see table 7.

TABLE 3

Station	Surface		Thermocline Beginning			P Gradient End		2000 Meter
	T°	P Mg/M³	Depth	T°	P Mg/M³	Depth	P Mg/M³	P Mg/M³
1208	25.20	3	100	24.18	3	795	36	28
1209	25.33	3	91	24.57	3	829	37	29
1210	23.95	2	100	21.46	2	1000	36	29
1211	24.10	1	84	22.06	2	872	30	(28)
1212	22.99	2	100	21.65	3	985	33	27
1213	23.40	2	90	21.90	1	969	34	24
1214	21.88	2	100	20.02	1	796	31	(25)
1215	22.18	2	200	19.55	2	958	32	25
1216	20.12	3	188	18.67	3	1039	32	24
1219	19.72	5	300	18.03	9	958	32	27

"Atlantis" stations between Haiti and Bermuda, April 1932. Thermocline beginning estimated as noted on page 16; 2000 meter values scaled from station curves; bracketed values estimated. End of phosphate gradient is estimated depth at which rapid increases of phosphate content, with increasing depth, cease; this is also the depth of phosphate maxima except for station 1211 where 31 mg P per cubic meter was recorded at 1063 meters. For station positions see table 7.

Between Bermuda and Chesapeake Bay[14]

In this region phosphate observations have been obtained at three different seasons: December 5–7, 1931 (stas. 1125–1130); February 12–15, 1932 (stas. 1134–1142); and April 17–23, 1932 (stas. 1220–1231). Since the April observations came near the end of a long cruise (standard phosphate solutions and reagents had not been renewed for nearly three months), the results may not be as reliable as the others but they are used in this discussion because this line of stations is unusually complete and illustrates better the essential features of vertical distribution of phosphate.[15] The general situation (fig. 12) differs from all others previously described in that, after extending about 400 miles westward from Bermuda, the section crosses the principle American coastwise thermal convergence,[16] where transitions between high and low values of temperature, oxygen, phosphate, etc. are more abrupt than in any other part of the area.

Surface. East of the convergence, surface observations indicate that from December (temperature 22.09°–22.50°) to February (temperature 19.38°–21.85°) phosphate content decreased from 4–5 to 0–3 mg P per cubic meter with no evidence of further change in April (0–2 mg P per cubic meter, temperature 18.85°–20.45°).

West of the convergence (to the continental shelf), seasonal variation of phosphate in the surface water is better defined, the records showing a decrease from 8–11 mg P in February to 2 mg P per cubic meter in April (tables 4, 5; figs. 12, 13). This reduction of phosphate occurs during the season at which plant growth presumably is accelerated and is depleting the surface layers of nutrient materials. But as the condition is also associated with a westerly displacement of the convergence during the same season it is possible that the reduction of phosphate content may not be due entirely to biological activity.

[14] Observed salinity range for the section was 34.90 o/oo to 36.67 o/oo; factor for correcting phosphate values for salt error is 1.35 (Brujewicz and Krasnova, 1933).

[15] Observations were also obtained in this region August 28–September 3, 1932 (stas. 1361–1373) but results are rejected because the cloudy phosphomolybdate blue colors prevented good color comparisons; results from stations 1143–1145 (February 15–18, 1932) are also discarded for the same reason.

[16] See Seiwell (1934), page 56.

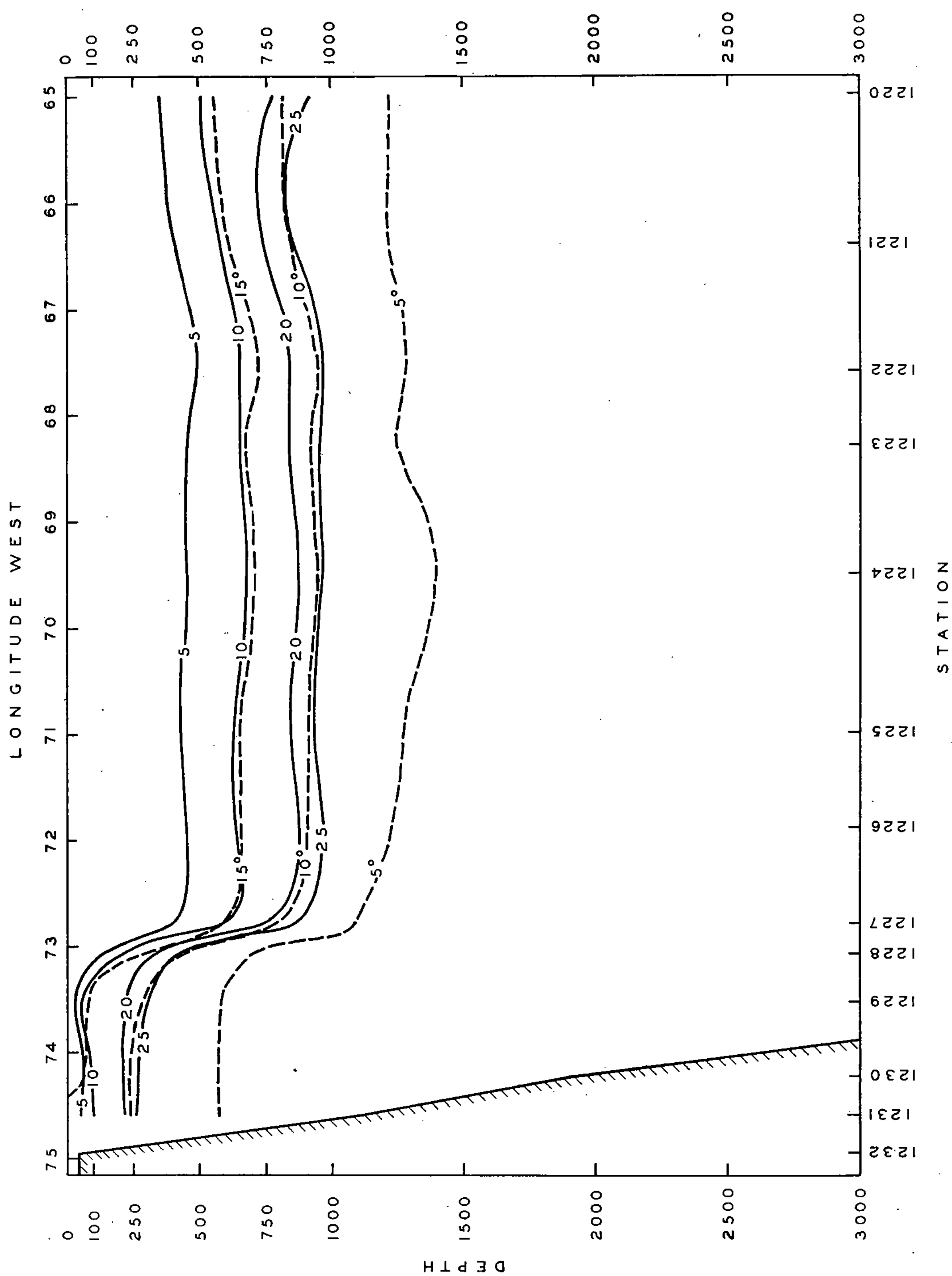

FIG. 12. Distribution of phosphate (mg P per m^3) between Bermuda and Chesapeake Bay, based on "Atlantis" stations 1220–1231; April 1932. For station positions see table 7.

The surface water overlying the convergence off Chesapeake Bay was distinguished from the water on either side by higher temperature both in February and April (table 5). In February the narrow band of warmer surface water (temperature 22.93°–24.05°) occurred between longitudes 71°51′W and 72°40′W (stas. 1138–1137), east and west of it surface temperatures were 21.85° and 12.05° respectively; the temperature drop to

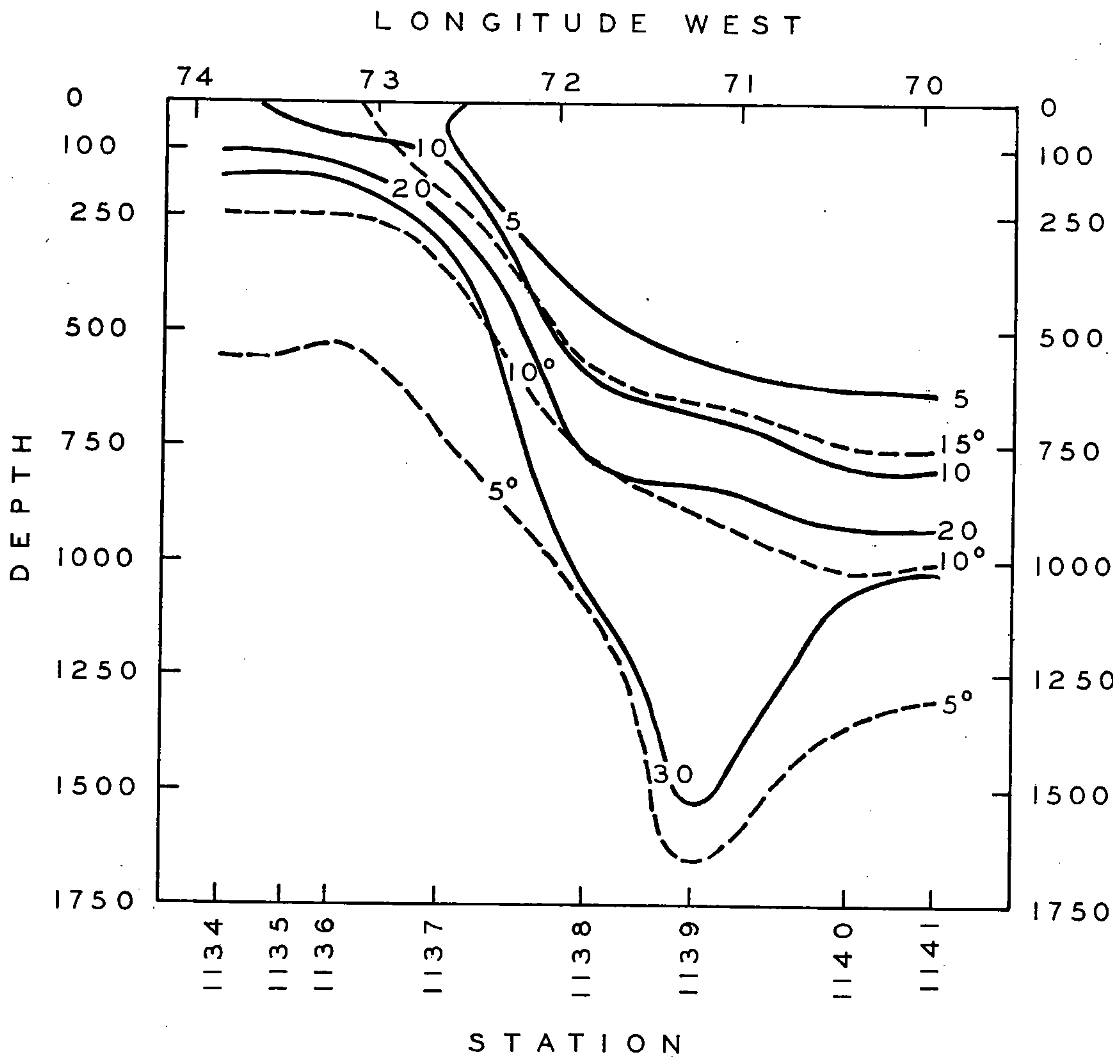

Fig. 13. Distribution of phosphate (mg P per m³) across principle American coastwise convergence between Chesapeake Bay and Bermuda, based on "Atlantis" stations 1134–1141; February 1932.

the east being 2.2° as compared to 10.9° to the west. Somewhat more than two months later the situation had changed so that the warmer band of surface water (22.11°–23.40°) occurred between longitudes 72°47′W and 73°05′W (stas. 1227–1228) showing a mean westerly displacement of about 40 minutes of longitude (equivalent to an average daily horizontal displacement of almost 36 seconds of longitude); the surface temperature drop, east and west of the warm band, being 3.4° and 4.6° respectively. Thus, while in April the temperature drop east of the convergence was one degree greater than in February the drop to the west had diminished more than 6 degrees, due to increased surface temperature of the slope water (between convergence and continental shelf). And as this water had not yet been much warmed by direct solar radiation[17] there appears to have been an influx of warmer, more saline, and phosphate poor water, which, by mixing and replacement, may account for the reduction of phosphate content during the interval.

Subsurface. Descending below the surface, figure 12 shows that the phosphate poor layer (less than 5 mg P per cubic meter) extends to depths of 400–500 meters east of the convergence, but only to 100 meters, or less, to the west of it. In still deeper water there is a similar distinction between vertical distribution on the two sides of the convergence; to the east phosphate content increased vertically to a maximum value of about 30 mg P per cubic meter in the vicinity of 1000 meters depth whereas to the west the maximum concentrations were found at depths of 300 to 500 meters. In the underlying water phosphate decreases but slightly so that maximum concentrations are not well defined, merely marking the limit at which rapid increases of phosphate cease with depth. The observations in December (1932) and February (1933) show somewhat higher phosphate values for the deep water than do those made in April, 1933 (figs. 12, 13, 14); in general, all observations from the section indicate that maximum phosphate values at intermediate depths may reach to about 35 mg P per cubic meter and in the still deeper water, e.g., at 2000 meters, may reach to about 29 mg P per cubic meter.

TABLE 4

Station	Surface		Thermocline Beginning			P Gradient End		2000 Meter
	T°	P Mg/M³	Depth	T°	P Mg/M³	Depth	P Mg/M³	P Mg/M³
1220	18.89	2	371	17.59	6	928	25	23
1221	19.97	1	364	17.99	2	949	30	23
1223	20.48	1	449	17.78	4	946	25	23
1224	19.15	2	415	18.15	2	993	28	24
1225	18.85	0	400	17.65	4	1000	31	24
1226	18.70	1	400	17.90	3	1070	27	24
1227	22.11	0	—	—	—	1010	28	24
1228	23.40	1	51	23.49	3	524	32	(23)
1229	18.78	2	36	17.92	7	426	24	21
1230	19.12	2	25	19.10	1	300	28	—

Stations between Bermuda and Chesapeake Bay, April 17–23, 1932. For positions see table 7.

Vertical phosphate distribution in the vicinity of the convergence is subjected to considerable variation as the latter is displaced horizontally. In February, when the convergence was farthest east (fig. 13), it was spread over a wider area (about 1 degree of longitude) than in April (about one-half degree of longitude) after it had shifted westward and tightened (fig. 12); although the condition indicated in figures 12 and 13 and table 5 may be somewhat exaggerated on account of the chance positions of the stations. To illustrate the tightening of the convergence we may use the course of the 10° isotherm (since all isolines of phosphate, oxygen, temperature, etc. follow the same general trend) which, in February, sloped upward from 880 meters depth at longitude 71°15′W to 260 meters at longitude 73°W (average decrease of depth being 5.9 meters per

[17] Station 1231, 36°36′N, 74°37′W, surface temperature was 11.36° on April 23, 1932. For discussion of temperature cycle on continental shelf of eastern United States see Bigelow (1933).

minute of longitude, fig. 13), whereas in April its course was from 880 meters at longitude 72°30′W to 270 meters at about longitude 73°40′W (average decrease of depth being 8.7 meters per minute of longitude, fig. 12). Thus, our data suggest that, with the westerly shift of the convergence, there is an intrusion of phosphate poor central At-

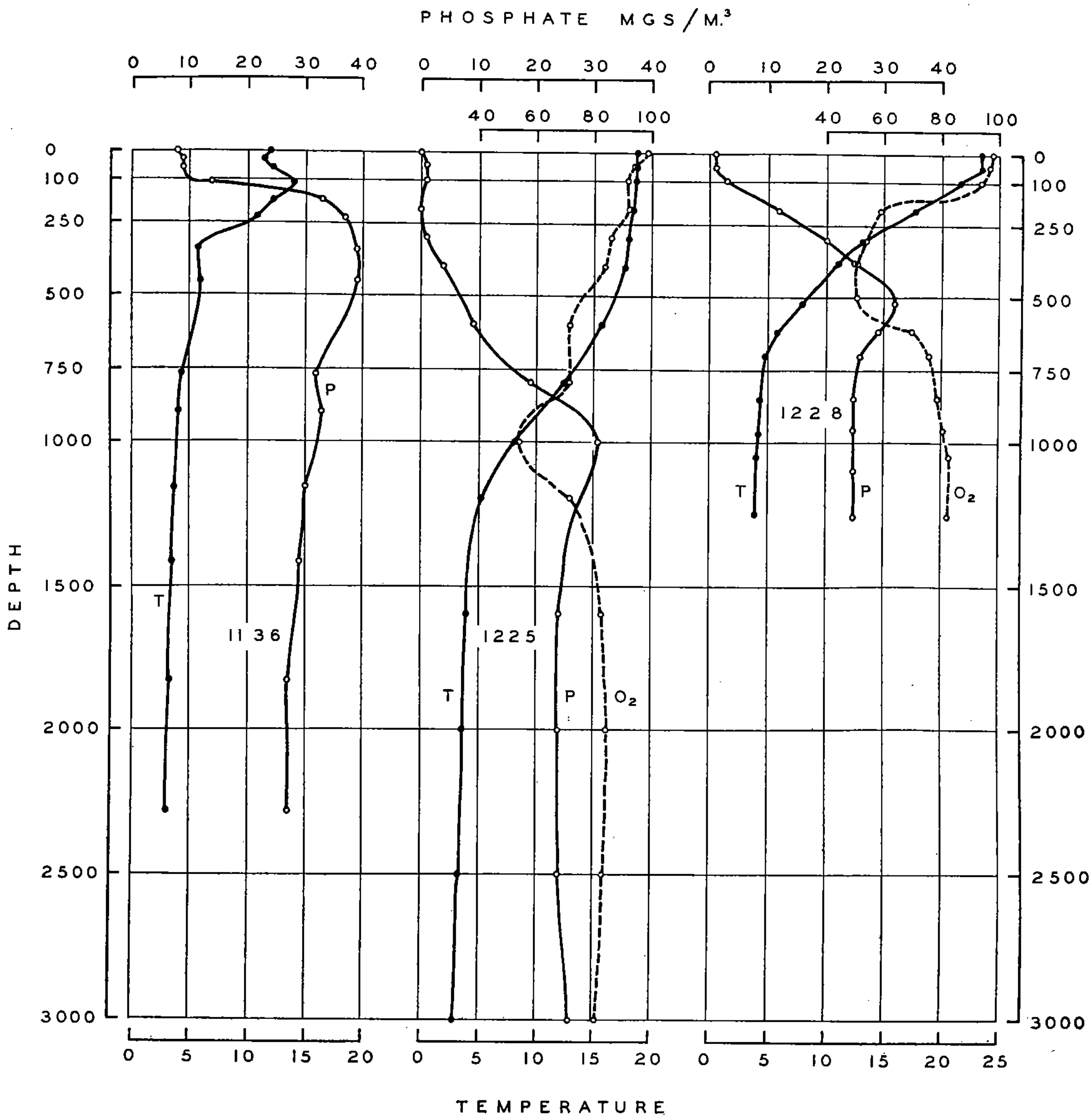

FIG. 14. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature between Bermuda and Chesapeake Bay; February and April 1932. For station positions see table 7.

lantic water which, through mixing, or otherwise, reduces the phosphate content of the water column between convergence and continental shelf. However, data are insufficient to estimate the possible effect of this condition on chemical fertility of the region or to postulate any further conclusions.

Surface temperatures along the section during April 17–23 (1932) had risen so little above the annual minimum (table 4) that water overlying the main thermocline still was close to its maximum homogeneity. East of the convergence, the principle thermocline is not sharply separated from the overlying water, and, in general, appears to begin at depths of 350 to 400 meters, whereas west of the convergence significant temperature decreases usually begin within 50 meters of the surface (fig. 15; table 4). In the water above the thermocline phosphate content is low and oxygen content near total saturation, the distribution of both elements being relatively uniform. The phosphate rich layers do not extend above the thermocline, and substances transported into this relatively homogeneous mass of water are distributed throughout it. West of the convergence the homogeneous layer may contain small, but distinct oxygen and phosphate gradients (stas. 1224, 1226; fig. 15) which are apparently below the depth of plant growth and at the time of observation illustrate temporary resistance of the layer to complete mixing; a resistance that gradually increases as the season advances (page 12). In general, characteristics of the water overlying the thermocline east of the convergence are similar to those for other parts of the north central part of the area while, west of the convergence, the relationships of phosphate, oxygen, and temperature were similar to those in the southern part of the area (page 13), although the vertical gradients are less strongly developed.

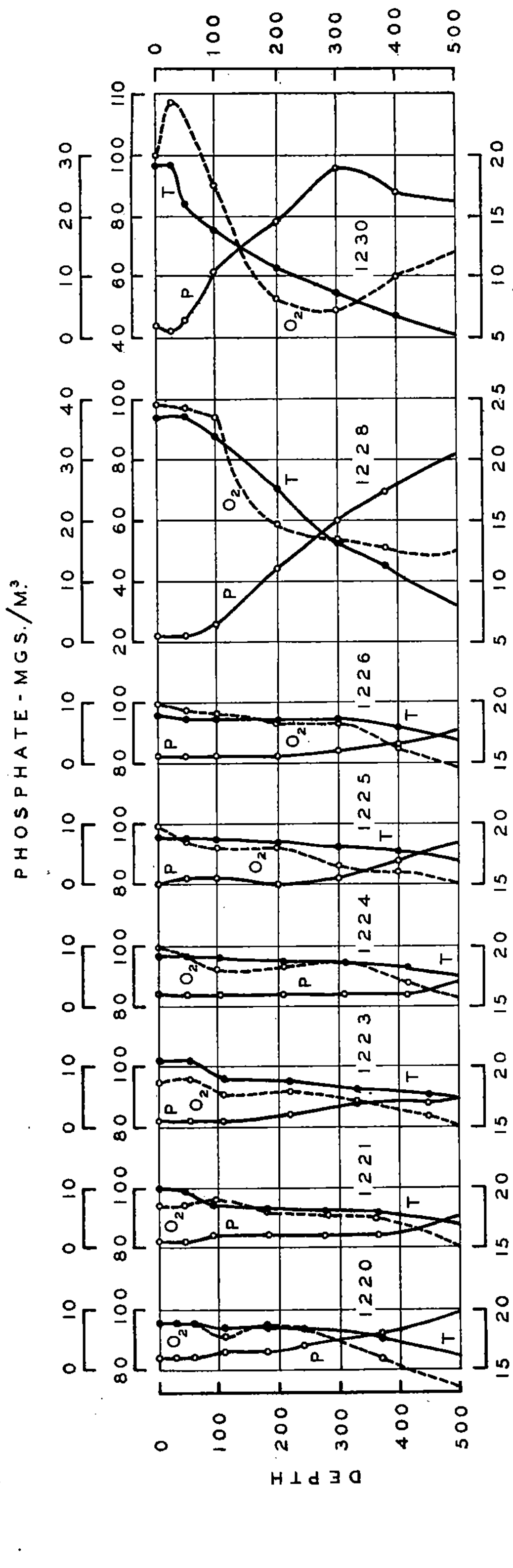

FIG. 15. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in the upper 500 meters between Bermuda and Chesapeake Bay; April 1932. For station positions see table 7.

TABLE 5

Station	West Longitude	Date 1932	Surface T°	Surface P Mg/M³	Surface S o/oo
1136	73°16′	Feb. 12	12.05	8	34.07
1229	73°32′	Apr. 22	18.78	2	36.27
1137	72°40′	Feb. 13	22.93	6	36.21
1138	71°51′	Feb. 13	24.05	2	36.30
1228	73°05′	Apr. 21	23.40	1	36.38
1227	72°47′	Apr. 21	22.11	0	36.40
1139	71°14′	Feb. 13	21.85	2	36.49
1226	71°53′	Apr. 21	18.70	1	36.63

Surface data for corresponding stations on both sides of the principal American coastwise convergence between Chesapeake Bay and Bermuda; February and April 1932.

Between Nova Scotia and Bermuda

Phosphate observations were obtained November 20–26, 1931 (stas. 1109–1124) and August 14–21, 1932 (stas. 1343–1357);[18] this discussion is based principally on the latter section[19] as the earlier results appear to be abnormally low for the three convergences off Nova Scotia.[20] There is no technical reason for doubting the verity of the November observations in this region because on the same cruise in water further south (using identical procedure and reagents), the vertical distribution of phosphate as determined then agrees well with that recorded nine months later (August, 1932) in the same region. However, the distribution of phosphate is sufficiently complicated without confusing the picture with doubtful observations so that we shall defer discussion of the November 1931 results from the three convergences (stas. 1107–1116) until such time as additional observations can be obtained for the region.

The vertical distribution of phosphate along the Nova Scotia Bermuda section during August 1932 (fig. 16) is similar to the Bermuda Chesapeake Bay section (fig. 12) in that about 300 miles north of Bermuda it crosses the principle American coastwise convergence, but differs from the latter section by crossing two additional convergences. Figure 16 illustrates that between Nova Scotia and Bermuda the principle American convergence (which is continuous from the vicinity of the Bahamas to the offing of Nova Scotia) was approximately 300 miles offshore in August 1932; the second about 200 miles and the third just off the continental shelf. The so-called second and third convergences may actually represent a large cyclonic vortex, developing on the inner boundary of the principle convergence and accumulating surface water in its own center.

[18] Two additional stations, 2136 (41°49′N, 65°26′W) and 2142 (41°12′N, 65°19′W), were obtained May 5 and 7, 1934.

[19] Observed salinity range was usually 35.0 o/oo to 36.5 o/oo; factor for correcting phosphate values for salt error is 1.35–1.36, see page 7.

[20] For a discussion of these convergences see Seiwell (1934). As a matter of convenience, the most easterly convergence extending from the Bahamas to the offing of Nova Scotia is called the first or principle convergence, and the two smaller convergences, between the first convergence and Nova Scotia are termed second and third convergences, respectively (fig. 16).

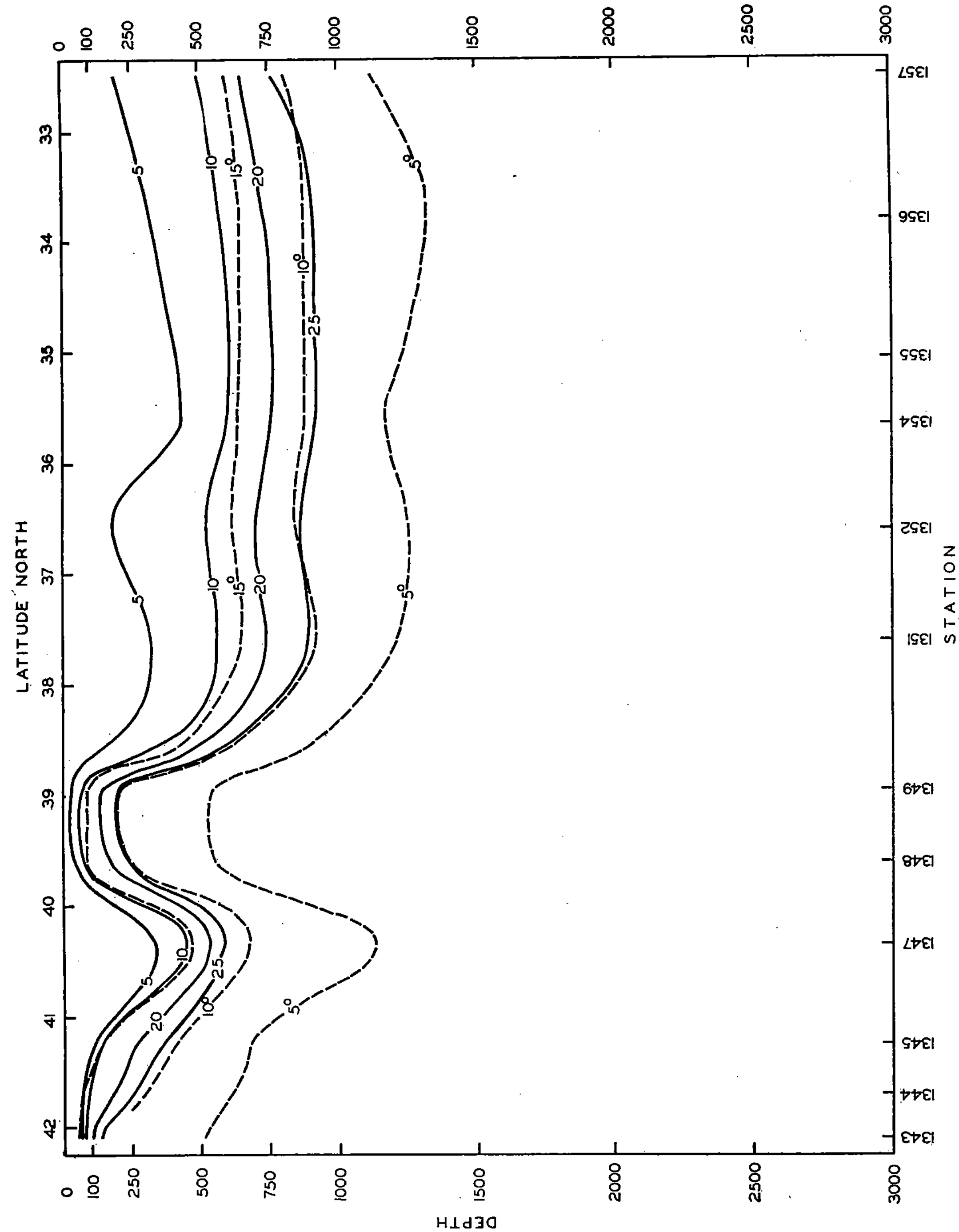

Fig. 16. Distribution of phosphate (mg P per m³) between Nova Scotia and Bermuda, based on "Atlantis" stations 1343–1357; August 1932. For station positions see table 7.

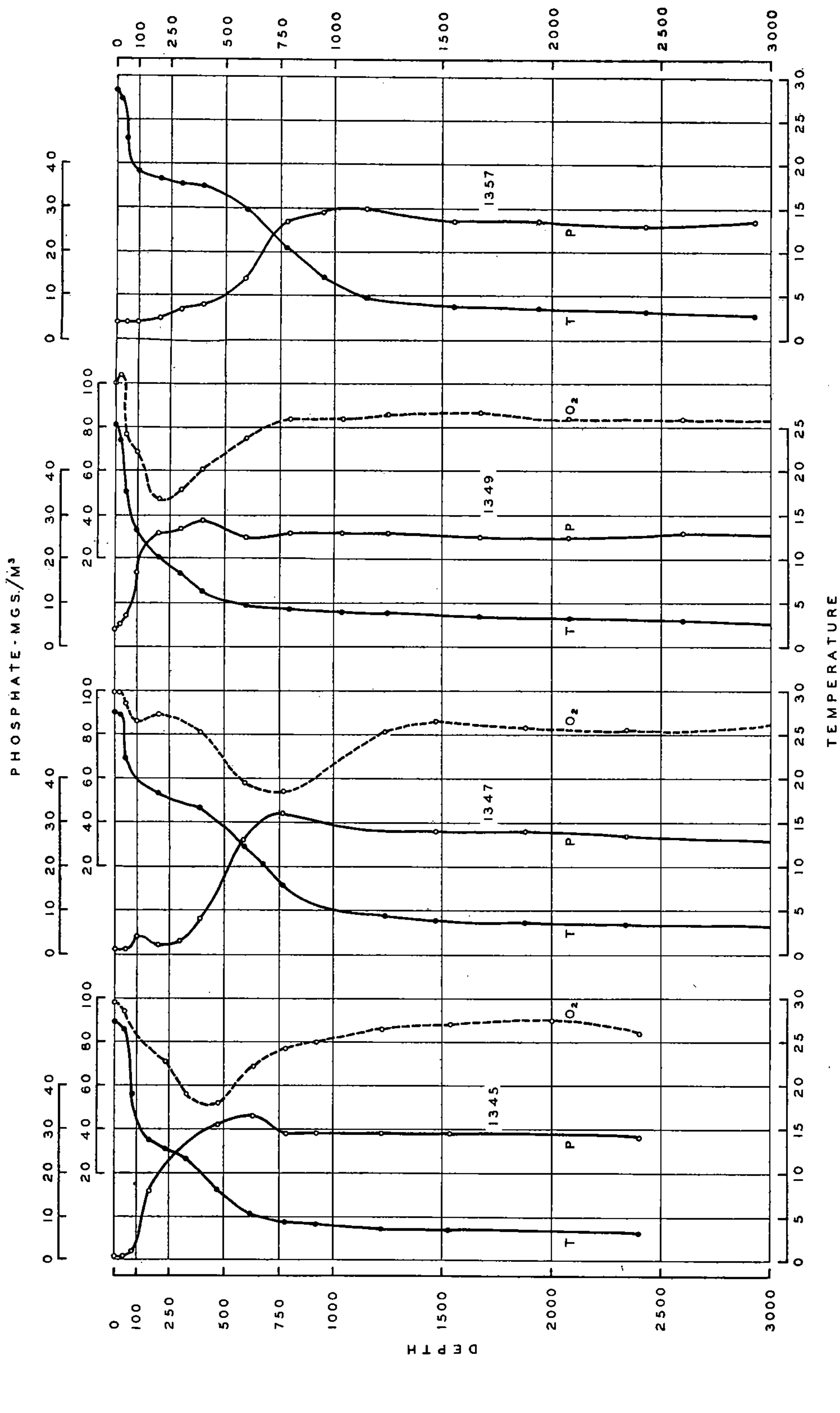

FIG. 17. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature between Nova Scotia and Bermuda; August 1932. For station positions see table 7.

Within all three convergences there are extreme transitions between high and low values of phosphate, oxygen, temperature, etc.; in going from south to north, all isolines slope sharply upward in the first convergence, downward in the second, and upward again in the third.

Comparison of surface observations shows that in November (surface temperature, 10.86°–25.21°) phosphate content ranged from 1 to 7 mg P per cubic meter and in August, nine months later (surface temperature, 16.88°–28.52°), from 1 to 4 mg P per cubic meter (table 6). This small variation is analagous to that recorded for the water between Bermuda and the convergence in the Bermuda Chesapeake Bay section (page 21).

Descending below the surface, figures 16, 17 and table 6 show that the phosphate poor layer, containing less than 5 mg P per cubic meter, extended to depths of 200–400 meters between Bermuda and the first convergence; to 30–40 meters between the first and second convergence; to about 340 meters between the second and third convergence; and then decreased again to about 100 meters after passing through the third or final convergence. In still deeper water, vertical distribution shows the same regional variation; as phosphate content rapidly increases with depth to 900 or 1000 meters between Bermuda and the first convergence, the vertical gradient ends at about 300 meters between first and second convergences; at about 800 meters between second and third convergences; and again at about 300 meters after passing through the third convergence. Phosphate content at the lower end of this gradient is the maximum, or nearly so, for the vertical water column, but is usually not sharply demarked from the underlying water (at 2000 meters P content is usually between 24 and 29 mg P/m^3).

Since vertical distribution of phosphate in the northern half of the section is regulated by the convergences, the situation as outlined for August 1932 (fig. 16) is not a permanent feature, but will change with the horizontal displacement of the convergences, evidence of which is given by comparing the courses of the 10° isotherms in November 1931 and August 1932. Thus, figure 18 shows that in November 1931 the slope of the isotherm within the first convergence occurred between 38° and 40°10′N latitude whereas in August 1932 it was between 37°30′ and 39°N. This southward shifting (30 minutes of latitude at its deeper end and about 70 minutes at its highest point) is most reasonably explained on the basis of a corresponding southerly displacement of the first convergence. A similar displacement is indicated for the second and third convergences, and between these latter convergences it is seen that in November the 10° isotherm only reached a depth of about 360 meters whereas in August it was observed at about 675 meters.

Hence, even though the picture may be somewhat distorted by the chance positions of the stations it is apparent that in the region dominated by the three convergences off Nova Scotia, significant changes may occur in the character of the water column; consequently it is probable that phosphate distribution will alter correspondingly at any given locality. And, as our knowledge of subsurface circulation is scanty, conclusions as to the probable limits of phosphate content for the deeper layers of the sea are, at best, tentative.

Surface temperature along the Bermuda Nova Scotia section in August was near the annual maximum so that the water overlying the main thermocline had developed its maximum stability as shown by the secondary or summer thermocline present in the uppermost layers (fig. 19). Within the convergences off Nova Scotia the depth at which the main thermocline began was in general indeterminate for the time of observation, although the distribution of phosphate and oxygen indicates that it began in the vicinity of 100 meters of the surface (stas. 1343, 1344, 1345; fig. 17), as is also the case in the

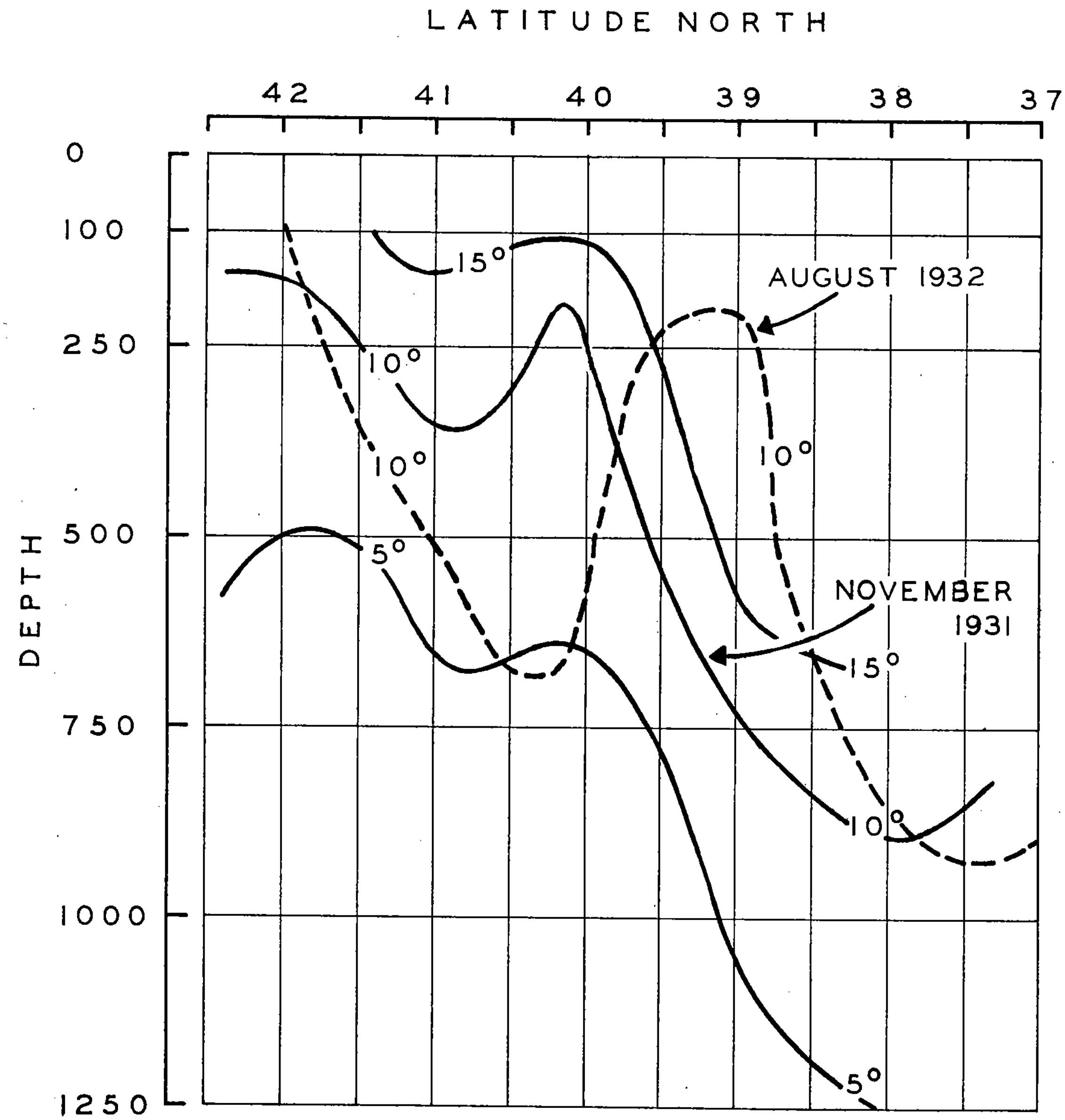

FIG. 18. Depth of 5°, 10° and 15° isotherms in three convergences between Nova Scotia and Bermuda in November 1931 and 10° isotherm in August 1932.

convergence off Chesapeake Bay (page 26). Between the second and third convergences (sta. 1347) where there is a thick accumulation of surface water, the main thermocline begins at about 400 meters depth. South of the first convergence the main thermocline

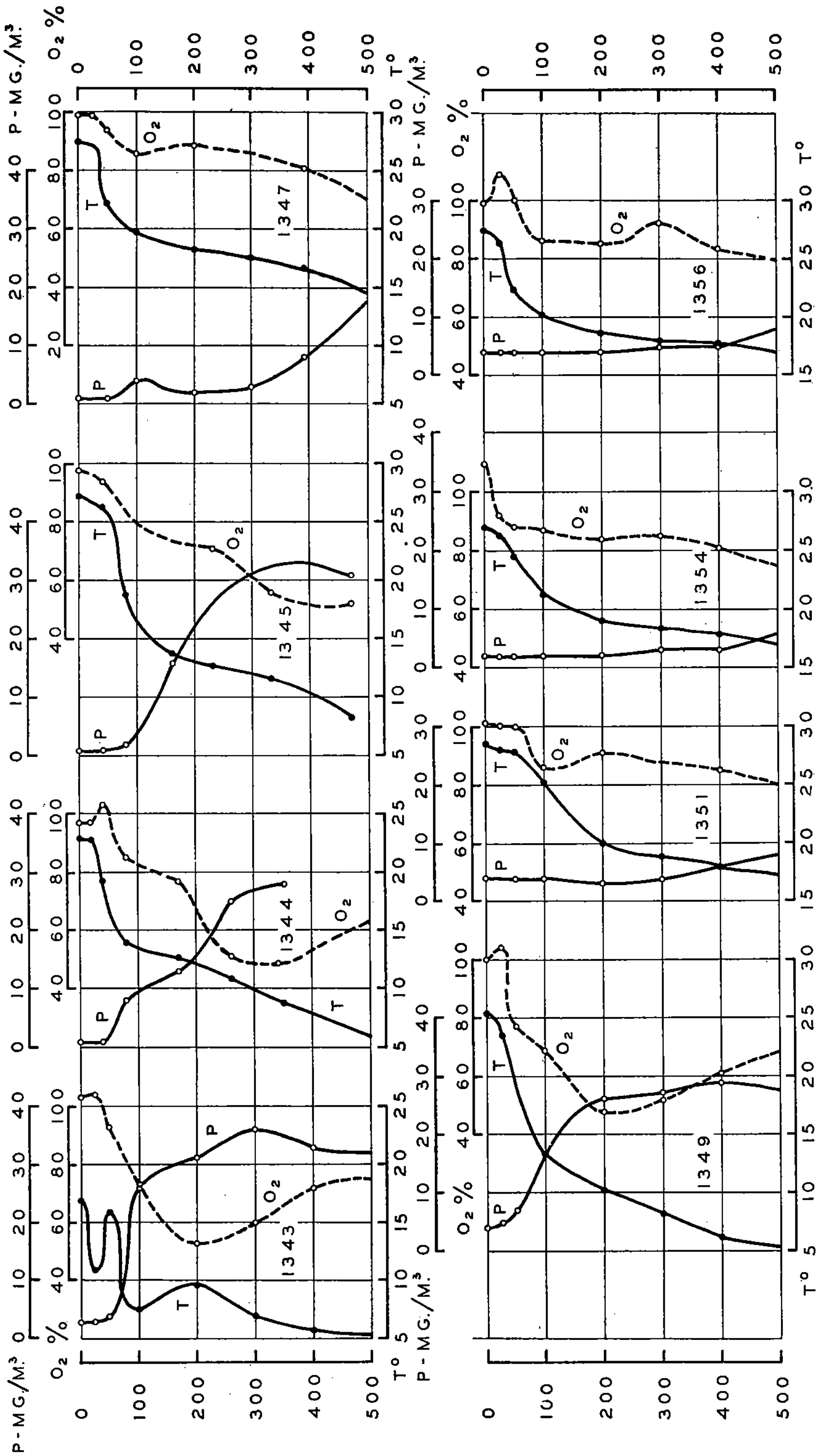

FIG. 19. Vertical distribution of phosphate (mg P per m³), oxygen (per cent of saturation), and temperature in the upper 500 meters between Nova Scotia and Bermuda; August 1932. For station positions see table 7.

TABLE 6

Station	Surface		Thermocline Beginning			P Gradient End		2000 Meter
	T°	P Mg/M³	Depth	T°	P Mg/M³	Depth	P Mg/M³	P Mg/M³
1343	16.88	3	—	—	—	300	36	26
1344	22.85	1	—	—	—	346	28	23
1345	27.20	1	—	—	—	628	33	29
1347	27.52	1	393	16.63	8	772	32	27
1348	26.86	2	—	—	—	289	27	—
1349	25.41	4	—	—	—	399	29	25
1351	28.42	4	400	18.04	6	983	26	24
1352	28.04	4	367	17.84	7	882	26	24
1354	26.89	2	400	17.85	3	1000	27	25
1356	27.41	4	400	17.71	5	943	26	26
1357	28.52	4	394	17.52	8	951	29	26

Stations between Nova Scotia and Bermuda, August 14–21, 1932. For station positions see table 7.

also begins at about 400 meters and the water overlying it is poor in phosphate and rich in oxygen (stas. 1345, 1356), similar to conditions in other parts of the north central portion of the area (page 9).

HORIZONTAL DISTRIBUTION OF PHOSPHATE PHOSPHORUS

Combination of "Atlantis" phosphate observations with those of the "Meteor" (Wattenberg, 1934) from the extreme southeastern portion of the area has allowed the construction of the accompanying charts of horizontal distribution (figs. 21 to 24). All stations from which data were obtained are entered in figure 1, and in order to illustrate possible magnitudes of deviation from the distribution as shown by the charts, some observed horizontal variations in the deep water of the North Atlantic may be briefly discussed.

Local Variations of Phosphate in the Subsurface Water of the North Atlantic (As Shown By Observational Data)

Observations between Chesapeake Bay and Bermuda (fig. 20), stations 1130 (34°36′N, 70°40′W, Dec. 7, 1931), 1140 (35°19′N, 70°24′W, Feb. 14, 1932), and 1225 (34°43′N, 71°00′W, April 20, 1932), show that phosphate content at any horizontal level below 1000 meters, in general, deviated from a mean value by about 2 mg P per cubic meter, within a radius of 60 miles. Also between Nova Scotia and Bermuda stations 1120 (37°53′N, 62°45′W, Nov. 24, 1931) and 1351 (37°37′N, 63°50′W, Aug. 18, 1932), about 50 miles apart, a similar although somewhat smaller variation occurred below 1000 meters. In general, the observed values in this region indicate that at mid depths the phosphate maxima is between 27 and 35 mg P per cubic meter and at 2000 meters the usual range is 24 to 29 mg.

In the southern half of the area, "Atlantis" station 1169 (16°22′N, 41°10′W, March 13, 1932) compared with "Meteor" station 284 (16°35′N, 42°00′W, March 24, 1927; Wattenberg, 1933) about 50 miles distant, shows significant differences in phosphate content above 1200 meters (30–42 mg P/m³ at 400 meters and 43–54 mg P/m³ at 1000 meters) but generally good agreement below 1400 meters depth. But "Atlantis" sta-

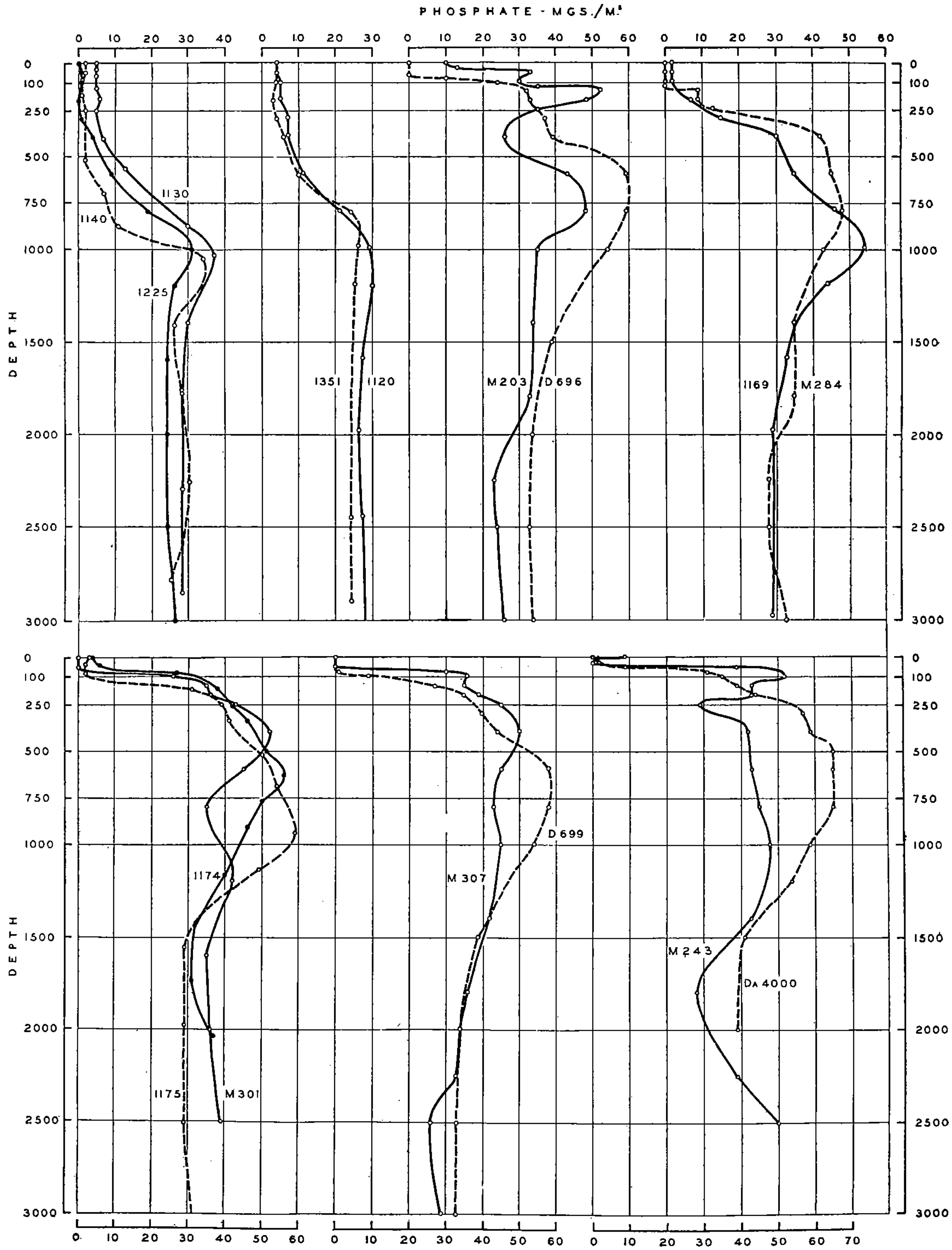

FIG. 20. Vertical distribution of phosphate (mg P per m³) in various localities as determined by "Atlantis" (unlettered stations), "Meteor" (M stations), "Dana" (Da stations), and "Discovery" (D stations).

tions 1174 (8°20′N, 40°45′W, March 16–17, 1932), 1175 (6°50′N, 40°25′W, March 17, 1932), still further south, compared with "Meteor" station 301 (8°07′N, 39°55′W, April 25–26, 1927; Wattenberg, 1933), all within an area of 50 miles radius show greater deviations; with horizontal range of 35 to 57 mg P per cubic meter at 800 meters and of 29 to 39 mg at 2500 meters depth.

In the eastern basin of the North Atlantic (fig. 20) observations at "Meteor" station 263 (8°12′N, 29°31′W, Feb. 12, 1927) differ by roughly 8–10 mg P throughout most of the water column (to 3000 meters depth) from those at "Discovery" station 696 (8°54′N, 30°02′W, May 12, 1931), about 50 miles distant. Further north "Meteor" station 307 (14°32′N, 29°38′W, May 2–3, 1927) and "Discovery" station 699 (14°27′N, 30°02′W, May 14, 1931), about 20 miles apart, show a maximum horizontal range of 43 to 58 mg P per cubic meter at 800 meters, identical values between 1400 and 2200 meters and a horizontal range of 26 to 33 mg P at 2500 meters.

A little south of the equator phosphate content at "Meteor" station 243 (1°18′S, 9°31′W, Dec. 29–30, 1926) and "Dana" station 4000 (0°45′S, 11°01′W, March 4, 1930), about 100 miles distant, diverges (at identical levels) about 20 mg P per cubic meter between 400 and 800 meters and about 8 mg P per cubic meter at 2000 meters depth.

The above examples are sufficient to show that the phosphate content of the water column, as determined by various observers at different times, does not remain constant. As might be expected the greater variations are recorded in the upper 1200 meters, but throughout the entire water column significant unexplained alterations are liable to occur. It is possible that some of these represent observational errors (page 7) but further elucidation is not possible at this time. Changes in phosphate content of the water column in the region influenced by the American coastwise convergences presents a special case as these appear to be associated with horizontal movements of the convergences themselves (pages 24 and 30).

The accompanying charts present a picture of average conditions at various horizontal levels and the use of a suitable isoline interval, e.g. 10 milligrams, has eliminated local variations and brought out the more permanent significant features of the horizontal distribution of phosphate.

Regional Variation of Phosphate

Phosphate values at the surface are usually between 0 and 3 mg P per cubic meter, and surface variations in the area of investigation are principally local and seasonal rather than regional.

100 meters. Figure 21 shows that phosphate content at this level is less than 5 mg P per cubic meter throughout the greater part of the area and is, no doubt, regulated by photosynthesis and frequent mixing of this not very stable water. Regional variations are confined to the extreme western and southeastern parts of the basin where phosphate content reaches to more than 20 mg P per cubic meter and where vertical mixing with the surface layers is restrained and phytoplanktonic activity may be restricted by reduced temperature.

250 meters. Phosphate content throughout the entire central part of the region is less than 10 mg P per cubic meter and well defined horizontal gradients are confined to the same regions as at 100 meters (fig. 22). The positions of the American coastwise convergences are clearly marked out in the northwestern part of the area where phosphate

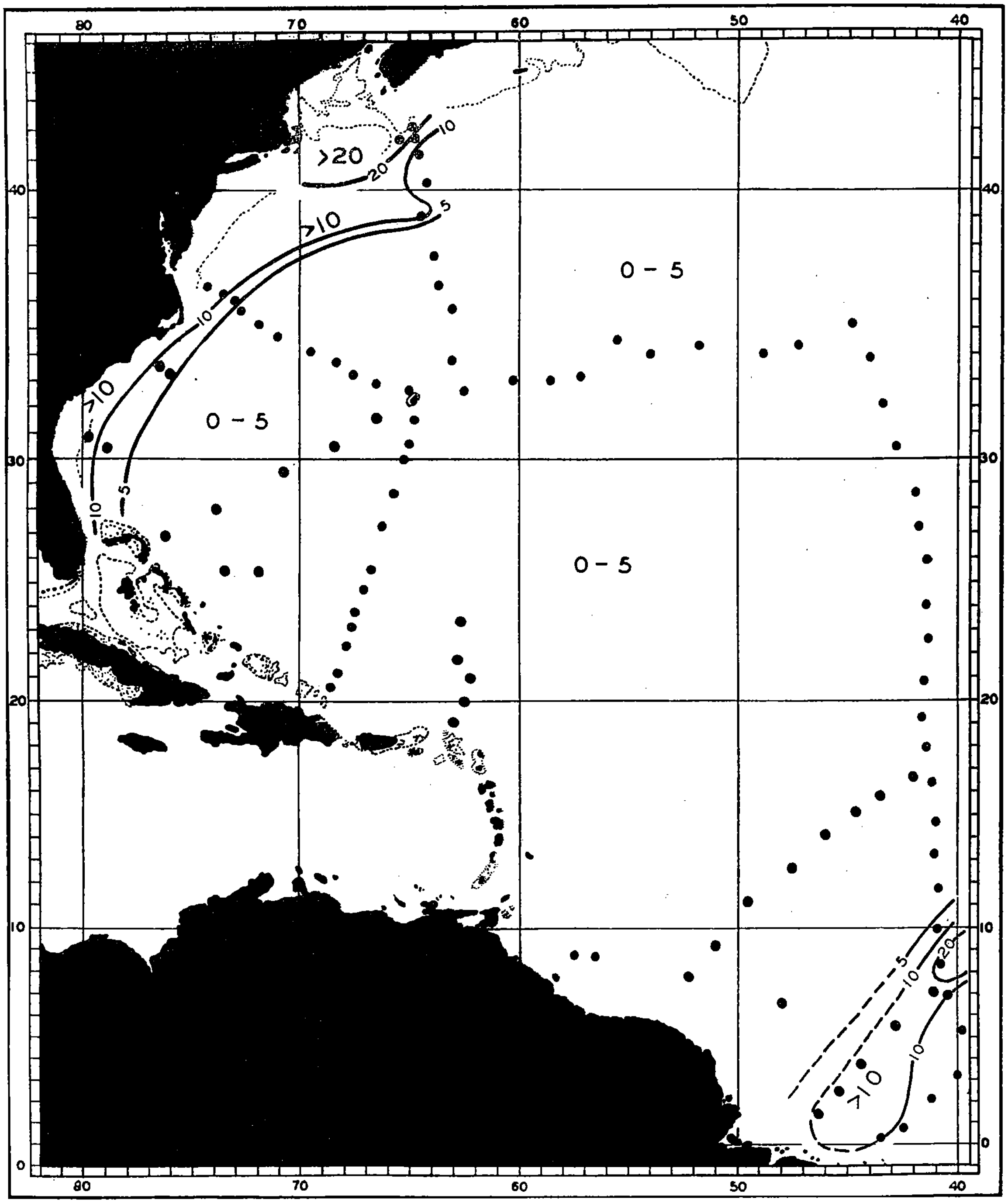

FIG. 21. Horizontal distribution of phosphate (mg P per m³) at 100 meters depth.

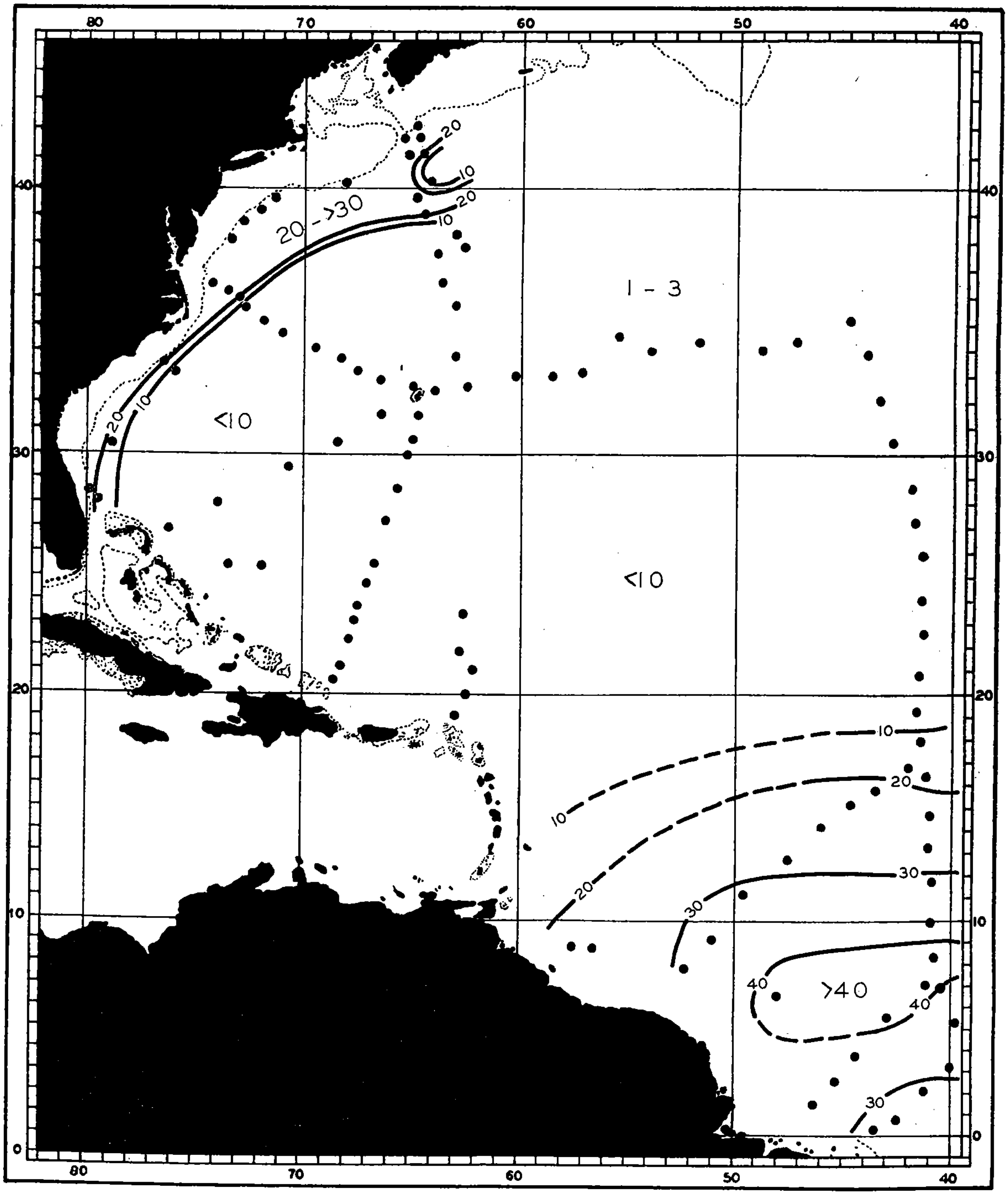

FIG. 22. Horizontal distribution of phosphate (mg P per m³) at 250 meters depth.

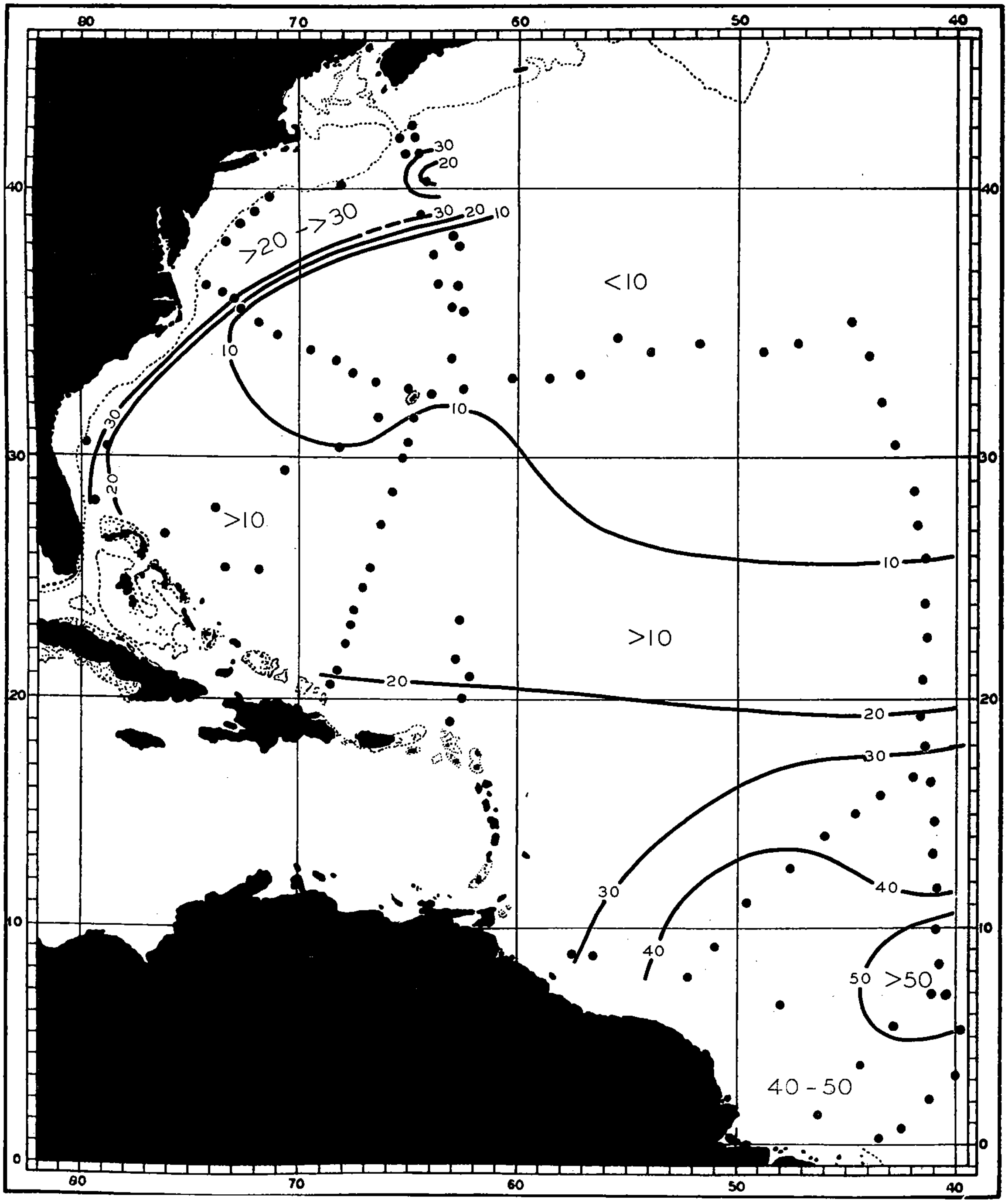

FIG. 23. Horizontal distribution of phosphate (mg P per m³) at 500 meters depth.

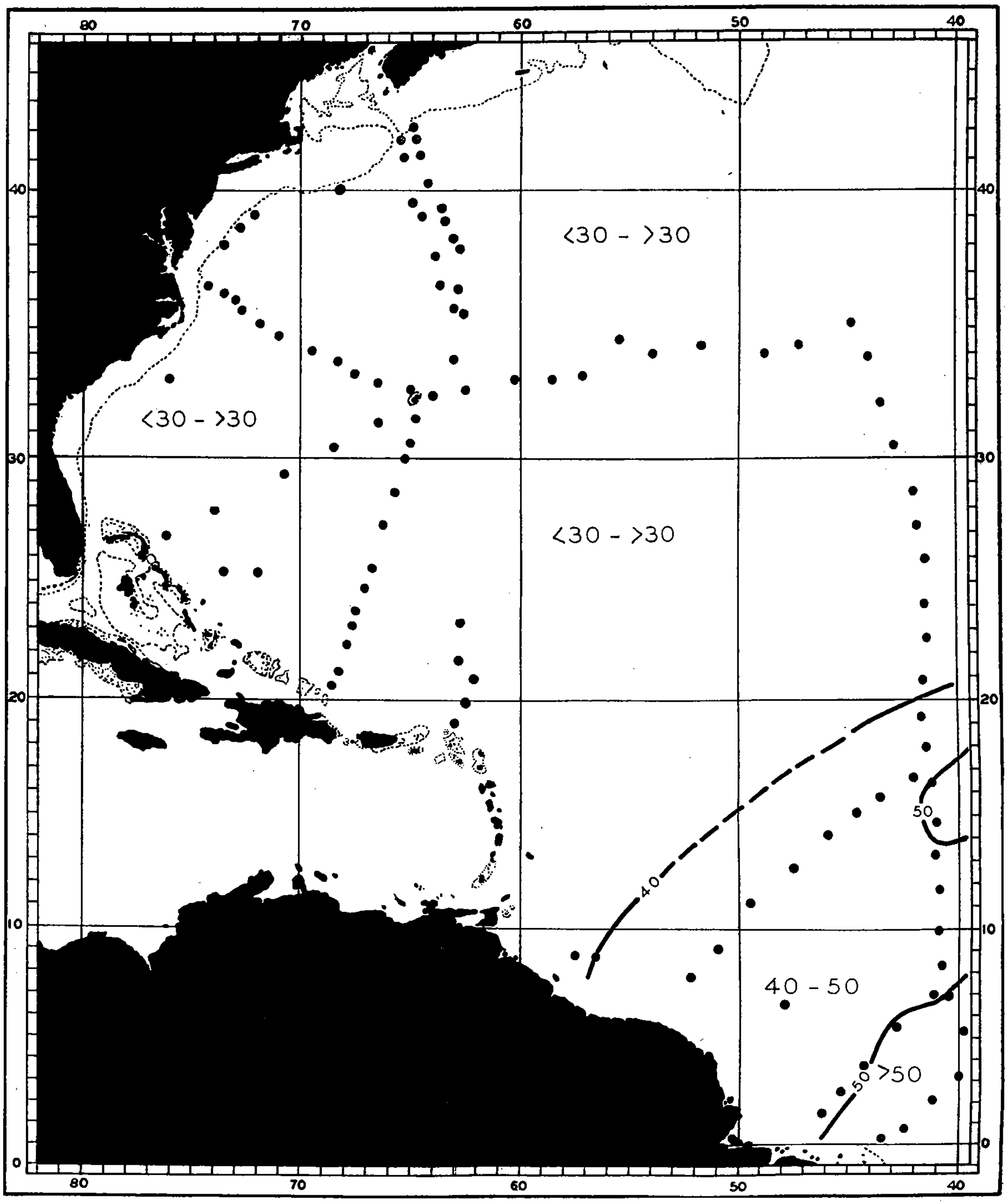

FIG. 24. Horizontal distribution of phosphate (mg P per m³) at 1000 meters depth.

concentration within a relatively short distance increases from less than 10 to more than 30 mg P per cubic meter (page 24). A similar gradient, although not so strongly developed, in the southeastern part of the region results from the north south increase in phosphate content (page 13); maximum concentrations of more than 40 mg P per cubic meter occur at about latitude 8°N.

500 meters. Although phosphate content is higher at this level, figure 23 shows that the general scheme of distribution is similar to that for 250 meters.

1000 meters. At this depth phosphate content is approaching a general uniformity characteristic of the deeper layers (fig. 24); north of latitude 25°N the concentration is usually close to 30 mg P per cubic meter[21] while to the south it increases to about 60 mg P per cubic meter at the extreme southeast end of the area.

In still deeper water horizontal variation of phosphate is relatively small; between 1000 and 3000 meters phosphate (P) concentration usually ranges from just under 30 to about 40 mg per cubic meter.

VERTICAL UPWARD TRANSPORT OF PHOSPHATE BY TURBULENCE

While nutrient substances are utilized apparently only in the illuminated surface layers of the sea, their ultimate vertical circulation is determined by the resultant physical and chemical qualities of the whole water column. It now being generally accepted that the deeper layers which contain relatively large quantities of phosphate act as a reservoir from which phosphate (and other nutrient substances) is withdrawn and transferred to the impoverished layer of plant activity, the question naturally arises to what extent are we able to interpret from the vertical distribution of phosphate and associated qualities in the area of investigation the probable significance of this vertical transport. The transporting agencies known to be significant in restoring nutrient substances to the surface layers of the sea are summarized by Gran (1932) as: (1) upwelling; (2) turbulence resulting from tidal currents; (3) vertical circulation set up by winter cooling of the surface layers, etc.; but they are significant principally in regulating the chemical fertility of coastal regions. In the offshore waters of the area of investigation an examination of the physical conditions and the distribution of phosphate suggest that, as the above agencies do not appear to be significant for phosphate transport, the most tangible factor likely to bring about vertical exchanges of phosphate is eddy conductivity.[22]

The Exchange Coefficient of Eddy Conductivity

The phenomenon of oceanic turbulence is generally believed to result from an impulse given to the water mass by currents, waves, etc., so that the water particles acquire irregular movements in many directions (Helland-Hansen, 1930). In the case of a horizontal current the water particles will also move up and down, and, in their vertical displacement from one level to another, carry with them their chemical and physical

[21] The abnormally low phosphate content of the deep water in the northern part of the north south section along the 40th meridian is an exception to this generalization; see page 13.

[22] The possibility must not be precluded that in the tropical Atlantic upwelling water may play a more or less significant part in supplying nutrient materials to the surface layers, as for instance in the region near latitude 8°N along the 40th meridian (figs. 6 and 7). However, from general conditions in this region it appears that any upwelling of subsurface water must occur at a rate so slow as not to greatly disturb the equilibrium of the surface water (see Seiwell, 1934).

properties. Thus, as turbulent motion brings about an exchange of properties between different levels, in the open ocean it may be a significant factor in determining regional chemical fertility.

The average vertical displacement of the water particles in a turbulent movement depends on the tangential or sheering stress (as caused by a laminar flow of sea water) and on the vertical stability of the water (as indicated by the vertical variation of density, σ_t). In general, great turbulence follows low vertical stability and high horizontal velocity, but the problem of the magnitude of eddy conductivity as related to the physical conditions in the sea is still an open question. Various investigators have studied the conditions and estimated the magnitude of the coefficient of eddy conductivity under certain specified conditions at particular times and places; and, in general, the results for ocean water are in agreement in that they indicate the numerical value of the eddy conductivity coefficient to be relatively small, usually less than 50 G. $Cm.^{-1}$ $Sec.^{-1}$.[23] Thus, Schmidt (1925) calculated from temperature and salinity variations eddy coefficients having magnitudes usually of 50 C.G.S. units or less for various coastal seas of northern Europe and for water off the Algerian coast.[24] Helland-Hansen (1930) reports values up to about 20 C.G.S. units for the uppermost water layers in the region off the European coast between 40° and 50°N latitude out to 20°W longitude. McEwen (1918) has obtained values of 30 to 40 C.G.S. units for the water off the California coast and Sverdrup (1933) calculated values for the conductivity coefficient of 20 to 50 C.G.S. units between depths of 200 and 500 meters in the not very stable Atlantic water ($E = 24 \times 10^{-8}$) north of latitude 80°N. In the greater depths of the ocean values of the eddy conductivity coefficient are even lower; Schmidt (1925) calculated, on the basis of the vertical temperature gradient, about 4 C.G.S. units for the deeper water of the Philippine Deep (5000 to 9800 meters); and my own calculation (Seiwell, 1934), on the basis of the disappearance of oxygen from the water, shows 8 to 9 C.G.S. units as characterizing the water of the tropical North Atlantic between 1000 and 2000 meters depth.

In considering the effect of eddy transfer on the vertical distribution of phosphate it is to be remembered that the magnitude of the transfer depends on the strength of the eddy conductivity coefficient and the vertical variation of phosphate, and that the direction of transfer is determined by the relative concentration of phosphate at the depths in question. Thus, if P be the concentration of phosphate at depth Z, and $P + \Delta P$ the concentration at $Z + \Delta Z$ the eddy transfer of phosphate will be from $Z + \Delta Z$ to depth Z, and the amount, N, transported per unit of time will be:

$$N = \frac{A}{\rho} \times \frac{dP}{dZ}$$

where ρ is the density of the sea water, which for practical purposes can usually be neglected in the calculation.

It is apparent that the upper and lower limits of the part of the water column through which the upward transport of phosphate by eddy conductivity is assumed to occur

[23] This differs from the coefficient of eddy viscosity in the sea which appears to have larger values. Helland-Hansen (1930) points out that values of about 200 C.G.S. units for the viscosity coefficient appear to be common in the sea, and that the ratio of the two coefficients is probably of the order of magnitude of 10^{-1} or possibly 10^{-2}.

[24] On the basis of current rotation with depth Schmidt (1925) calculated an eddy coefficient of 250 C.G.S. units for the Florida stream.

must be defined on some rational basis. In doing this, in order that the significance of the analysis may not be obscured, we are governed by the following criteria: first, the part of the water column selected for analysis must be sufficiently stable so that significant quantities of phosphate are not likely to be removed except by eddy transfer, and, second, the overlying water must be less stable so that a more rapid distribution occurs throughout its entire thickness (by winds, surface cooling, etc.) and the transferred phosphate is quickly made available for plant utilization. Thus, for the present analysis, the beginning of the main thermocline (see footnote 7, page 9) has been selected as the depth of the upper boundary of the layer in which phosphate transfer by eddy conductivity is to be studied, since, in general, at this depth the water column permanently becomes more resistant to vertical circulation; the overlying water may be homogeneous during a large part of the year. The position of the main thermocline in the water column is discussed on page 16, and its relation to the biological consumption of phosphate on page 50. The previous discussion of phosphate and oxygen content of the homogeneous layer suggests that biological activity involving phosphate consumption takes place principally above the thermocline, so that little, if any, direct consumption of phosphate would occur in the more stable underlying water.

The lower limit of the upward transfer of phosphate is naturally marked out by the depth at which significant increases of phosphate cease with increasing depth. This is, in general, the depth of the maximum phosphate concentration, discussed on page 9. The part of the water column thus included is, for convenience, termed the transport layer.

Position of Transport Layer In Water Column

The depth of the upper boundary of the transport layer, based on the beginning of the thermocline, varies from more than 400 meters to less than 50 meters within the area of investigation, according to the general picture of regional variation given in figure 8. The overlying water which may, at times, be completely homogeneous (page 16) possesses a high oxygen and low phosphate content, and, in general, definite decreases of oxygen and increases of phosphate begin at about the same depth as the decline of temperature. Since the depth at which the thermocline begins is frequently not well defined, the depths tabulated in tables 1 to 6 have been determined after detailed study of the various stations and it is possible that, for stations having a thick homogeneous layer, the error of estimation may sometimes be 50–100 meters.[25]

The depth of the lower boundary of the transport layer, marked out by the depth in the midstrata at which significant increases of phosphate cease with increasing depth, has also been discussed for each section. In practically all cases this phosphate gradient ends where the phosphate content for the vertical water column is at its maximum and the direction of eddy transfer is clearly illustrated by the vertical distribution curves in figure 7. The depths and concentrations determined for the lower end of the phosphate discontinuity layer are naturally not exact for individual stations because of the chance positions of the observations. Regional variations of depth and concentrations are illustrated by table 7; north of 20°N (except in the region influenced by the convergences) the phosphate gradient ends at about 800 to 1000 meters with a value of 38–49 mg P per cubic meter (corrected for salt error; see page 7). And south of 20°N, its depth decreases to about 600 meters (between latitude 5° and 10°N) with a value usually be-

[25] In doubtful cases the phosphate and oxygen gradients have been taken as guides (page 20).

tween 65 and 80 mg P per cubic meter (corrected for salt error). For reasons discussed later that part of the area influenced by the American coastwise convergences is omitted from this discussion.

Vertical Variation of σ_t and Phosphate Within the Transport Layer

Calculation of the regional vertical variation of σ_t makes possible a comparison of regional differences in the vertical stability of the transport layer;[26] values for individual stations are given in table 7. Vertical variations of σ_t per centimeter are smallest (1.4×10^{-5} to 1.8×10^{-5}) for the transport layer north of 25°N (east of the American convergences), greater to the south of latitude 25°N; the maximum vertical variation of 5.4×10^{-5} σ_t units per meter was found in the vicinity of latitude 10°N. Thus, in general, the stability of the transport layer is only about one half to one third as high in the northern half of the area as in the southern half. If this regional variation of stability be used as an approximate indicator of the relative magnitude of eddy conductivity existing within the transport layer, then in the northern part of the area (exclusive of the American coastwise convergences) the exchange coefficient will have approximately two to three times the magnitude of that in the southern part. In making this regional distinction certain conditions are presupposed which at present cannot be supported with data, e.g., that eddy conductivity is approximately inversely proportional to stability,[27] and that sheering stresses in this layer are relatively uniform throughout, but from general knowledge these assumptions appear reasonable enough to be used as a working hypothesis in the following discussion. It must be noted, however, that an obvious exception to a general relationship between eddy conductivity, stability and sheer, within the investigated area, is presented by conditions in the American coastwise convergences, where, high current velocities in the upper part of the water column (up to 100 cms/sec) no doubt produce greater sheering stresses for the given stability than in other parts of the area. Hence, since it is not practical to approximate the magnitude of eddy conductivity on the basis of stability variation, further discussion of the convergence region is omitted.

The vertical variation of phosphate in the transport layer is expressed here as the change of concentration (grams of P per cm³) per linear centimeter. Thus, if P be the phosphate content at depth Z (upper boundary), and $P + \Delta P$ that at depth $Z + \Delta Z$ (lower boundary), then:[28]

$$\text{Vertical variation of phosphate} = \frac{\Delta P}{\Delta Z} \text{ (grams/cm}^3\text{/cm)}.$$

[26] If σ_t be the density at depth Z (meters), the upper limit of the transport layer; and $\sigma_t + \Delta\sigma_t$ that at $Z + \Delta Z$ (meters), the lower limit of the transport layer (depth of phosphate maxima), then:

$$\text{Vertical variation of } \sigma_t \text{ per meter} = \frac{\Delta\sigma_t}{\Delta Z}$$

For comparison of stability conditions in different parts of the region it is sufficient to consider the change with depth of $\sigma_t (\sigma_t = [S^t - 1]\ 1000$; where S^t is the density of sea water; Knudsen, 1901). A stability in conventional units of $E = 1000 \times 10^{-8}$ is almost equal to a vertical variation of 0.01 units of σ_t per meter. For a more exact calculation of stability see Hesselberg and Sverdrup (1915).

[27] In the light of general knowledge and information from Professor C. G. Rossby this assumption is reasonable for high or moderately high stabilities such as discussed in this paper, but is not valid for very low stabilities in which case eddy conductivity would approach infinity.

[28] P should be expressed as grams of phosphate per gram of sea water to be dimensionally correct.

TABLE 7

Station	Lat. North	Long. West	Upper Boundary Depth	Upper Boundary P Mg/M³	Upper Boundary σ_t	Lower Boundary Depth	Lower Boundary P Mg/M³	Lower Boundary σ_t	Thickness ΔZ	$\frac{\Delta\sigma_t}{\Delta Z(\text{cms})}$	$\frac{\Delta P(\text{g/cm}^3)}{\Delta Z\ (\text{cm})}$	A g cm^{-1} sec^{-1}
1147	32°37′	62°35′	382	4	26.49	808	45	27.24	426	1.76×10^{-5}	9.62×10^{-13}	26.9
1149	32°59′	60°16′	384	4	26.52	1002	42	27.63	618	1.79×10^{-5}	6.14×10^{-13}	26.4
1150	33°02′	58°37′	496	5	26.66	937	36	27.42	441	1.72×10^{-5}	7.03×10^{-13}	27.5
1151	33°11′	57°07′	400	4	26.51	1000	42	27.49	600	1.63×10^{-5}	6.33×10^{-13}	29.0
1152	34°48′	55°30	262	3	26.46	821	38	27.28	559	1.47×10^{-5}	6.25×10^{-13}	32.2
1153	34°02′	54°05′	363	3	26.46	836	41	27.32	473	1.82×10^{-5}	8.04×10^{-13}	26.0
1154	34°20′	51°45′	435	3	26.63	934	36	27.41	499	1.56×10^{-5}	6.61×10^{-13}	30.3
1155	34°02′	48°50′	285	3	26.51	892	42	27.53	607	1.68×10^{-5}	6.42×10^{-13}	28.2
1156	34°23′	47°11′	296	4	26.64	758	41	27.37	462	1.58×10^{-5}	8.00×10^{-13}	29.9
1157	35°10′	44°40′	351	4	26.55	1003	24	27.46	652	1.4×10^{-5}	3.07×10^{-13}	33.8
1158	33°52′	44°03′	281	2	26.53	752	26	27.20	471	1.42×10^{-5}	5.52×10^{-13}	33.3
1159	32°09′	43°30′	208	5	26.55	732	35	27.36	524	1.54×10^{-5}	5.72×10^{-13}	30.7
1160	30°26′	42°53′	154	2	26.18	954	31	27.53	800	1.69×10^{-5}	3.63×10^{-13}	28.0
1161	28°38′	41°57′	74	2	26.04	931	34	27.52	857	1.73×10^{-5}	3.73×10^{-13}	27.3
1162	27°15′	41°45′	100	1	25.91	995	36	27.55	895	1.83×10^{-5}	3.91×10^{-13}	25.9
1163	25°53′	41°25′	92	1	25.99	967	46	27.48	875	1.70×10^{-5}	5.14×10^{-13}	27.8
1164	24°06′	41°26	95	1	25.87	990	53	27.52	895	1.84×10^{-5}	5.81×10^{-13}	25.7
1165	22°35′	41°19′	99	1	25.69	795	44	27.40	696	2.45×10^{-5}	6.18×10^{-13}	19.3
1166	20°50′	41°38′	75	1	25.30	1024	50	27.57	949	2.39×10^{-5}	5.16×10^{-13}	19.8
1167	19°17′	41°40′	93	1	25.24	882	61	27.47	789	2.82×10^{-5}	7.60×10^{-13}	16.8
1168	17°55′	41°30′	94	1	25.22	759	65	27.39	665	3.26×10^{-5}	9.62×10^{-13}	14.5
1169	16°22′	41°10′	100	2	25.48	993	73	27.51	893	2.27×10^{-5}	7.95×10^{-13}	20.8
1170	14°47′	40°58′	97	1	25.01	784	73	27.37	687	3.43×10^{-5}	10.4×10^{-13}	13.8
1171	13°15′	41°06′	80	0	24.32	778	66	27.36	698	4.35×10^{-5}	9.45×10^{-13}	10.9
1172	11°43′	40°54′	76	0	24.10	662	65	27.25	586	5.37×10^{-5}	11.1×10^{-13}	8.8
1173	9°57′	40°55′	40	1	24.06	629	73	27.25	589	5.41×10^{-5}	12.2×10^{-13}	8.8
1174	8°20′	40°45′	0–43	5	24.08	634	76	27.27	634	5.01×10^{-5}	11.2×10^{-13}	9.4
1175	6°50′	40°25′	84	2	24.27	691	76	27.30	607	5.00×10^{-5}	12.2×10^{-13}	9.5
1176	5°16′	39°47′	85	4	23.43	852	80	27.43	767	5.21×10^{-5}	9.9×10^{-13}	9.1
1177	3°13′	40°00′	53	2	23.66	817	82	27.44	764	4.95×10^{-5}	10.5×10^{-13}	9.6
1179	0°45′	42°38′	47	2	23.55	890	73	27.39	843	4.56×10^{-5}	8.42×10^{-13}	10.4
1208	20°38′	68°36′	100	4	24.88	795	49	27.35	695	3.55×10^{-5}	6.47×10^{-13}	13.3
1209	21°19′	68°13′	91	4	24.61	829	51	27.54	738	3.97×10^{-5}	6.40×10^{-13}	11.9
1210	22°14′	67°50′	100	3	25.75	1000	49	27.57	900	2.02×10^{-5}	5.11×10^{-13}	23.4
1211	23°10′	67°34′	84	3	25.53	872	41	27.38	788	2.35×10^{-5}	4.82×10^{-13}	20.1
1212	23°46′	67°24′	100	4	25.63	985	45	27.51	885	2.12×10^{-5}	4.63×10^{-13}	22.3
1213	24°45′	67°05′	90	1	25.56	969	46	27.49	879	2.20×10^{-5}	5.12×10^{-13}	21.5
1214	25°33′	66°45′	100	1	26.04	796	42	27.29	696	1.80×10^{-5}	5.88×10^{-13}	26.3
1215	27°12′	66°11′	200	3	26.14	958	43	27.40	758	1.66×10^{-5}	5.27×10^{-13}	28.5
1216	28°33′	65°43′	188	4	26.33	1039	43	27.53	851	1.42×10^{-5}	4.58×10^{-13}	33.3
1219	31°30′	64°31′	300	12	26.45	958	43	27.49	658	1.58×10^{-5}	4.7×10^{-13}	29.9
1220	32°40′	65°00′	371	8	26.51	928	34	27.48	557	1.74×10^{-5}	4.67×10^{-13}	27.2
1221	32°51′	66°25′	364	2	26.45	949	41	27.38	585	1.59×10^{-5}	6.67×10^{-13}	29.7
1223	33°41′	68°18′	449	5	26.48	946	34	27.26	497	1.57×10^{-5}	5.83×10^{-13}	30.1
1224	34°10′	69°31′	415	2	26.44	993	38	27.23	578	1.37×10^{-5}	6.23×10^{-13}	34.5
1225	34°43′	71°00′	400	5	26.49	1000	41	27.38	600	1.48×10^{-5}	6.00×10^{-13}	31.9
1226	35°07′	71°53′	400	4	26.46	1070	36	27.55	670	1.63×10^{-5}	4.78×10^{-13}	29.0
1228	35°57′	73°05′	51	4	24.72	524	43	27.35	473	5.56×10^{-5}	8.25×10^{-13}	—
1229	36°12′	73°32′	36	9	26.16	426	32	27.48	390	3.38×10^{-5}	5.90×10^{-13}	—
1230	36°27′	74°15′	25	1	25.91	300	38	27.36	275	5.27×10^{-5}	13.5×10^{-13}	—
1347	40°15′	64°10′	393	11	26.62	772	43	27.39	379	2.03×10^{-5}	8.44×10^{-13}	—
1351	37°37′	63°50′	400	8	26.43	983	35	27.42	583	1.70×10^{-5}	4.6×10^{-13}	27.8
1352	36°36′	63°37′	367	10	26.46	882	35	27.29	515	1.61×10^{-5}	4.85×10^{-13}	29.4
1354	35°40′	63°00′	400	4	26.46	1000	37	27.44	600	1.63×10^{-5}	5.5×10^{-13}	29.0
1356	33°47′	63°09′	400	7	26.48	943	35	27.38	543	1.66×10^{-5}	5.2×10^{-13}	28.5
1357	32°30′	63°57′	394	11	26.54	951	39	27.50	557	1.72×10^{-5}	5.0×10^{-13}	27.5

Data for transport layer (page 42) in area of investigation. Phosphate values corrected for salt error (page 7); for calculation of vertical variation of $\sigma_t\left(\frac{\Delta\sigma_t}{\Delta Z(\text{cms})}\right)$ see page 43; of vertical variation of phosphate $\left(\frac{\Delta P}{\Delta Z}\ \frac{\text{g/cm}^3}{\text{cm}}\right)$ see page 43; and of eddy conductivity coefficient (A) see page 45.

The results for individual stations (table 7) with the average results for 5° coordinate squares (fig. 25) show that throughout the whole western basin the average vertical variation of phosphate within the transport layer is usually between 4×10^{-13} and 11×10^{-13} grams/cm^3/cm. Maximum phosphate gradients occur in the southern part of the region corresponding to the high phosphate concentrations in the midstrata (up to 80 mg/m^3; corrected for salt error). North of latitude 25°N vertical variation of phosphate is usually between 5×10^{-13} and 7×10^{-13} g/cm^3/cm; the relatively low values recorded for squares 11, 12, and 13 are apparently due to the low phosphate concentration at mid depths in this region (see page 13).

Estimation of Regional Variation of Coefficient of Eddy Conductivity

If the vertical variation of σ_t be used now to estimate regional variation of eddy conductivity, as brought out in the previous discussion these results may in turn be combined with vertical variation of phosphate to approximate the eddy transfer of that substance through the transfer layer, e.g., from phosphate rich midstrata to phosphate poor overlying water.

Data on the probable theoretical relationship of eddy conductivity and stability were derived from relationships existing in the tropical western North Atlantic; in the region between latitudes 13°N and 3°N along the 40th meridian. In this region it has been estimated that the annual phosphate consumption is 75.6×10^{-5} grams P per square centimeter of surface in the warm surface layer overlying the thermocline[29] (less than 100 meters thick; figs. 6, 7) and also that the presence of this nutrient substance is due both to eddy transfer from the phosphate rich midstrata and to regeneration of phosphate within the photosynthetic layer itself.[30] Proceeding from this, we may next estimate the amount of phosphate enrichment resulting from eddy transfer.

Figure 8 shows that in the tropical North Atlantic the upper part of the thermocline (also the upper part of the transport layer, but not the whole of it) is an extremely well developed discontinuity layer, less than 100 meters thick, which probably represents one of the most extreme cases of stability for any water layer of similar thickness in the North Atlantic basin (compare stability data below with that of Helland-Hansen, 1930). And, it it likely that the eddy coefficient which can exist in it is very small, perhaps not greater than 2 C.G.S. units (see page 40). Hence, we have first calculated the average annual phosphate transport through this narrow discontinuity layer, which would result from its phosphate gradient[31] (table 8; fig. 8) and an assumed eddy coefficient of 2 C.G.S. units. Then, assuming the result to be equal to the average total eddy transfer of phosphate through the entire transport layer of the same region, it is possible, by reversing the procedure, to calculate from the phosphate gradient (of the transport layer; table 7) an average eddy coefficient for the whole transport layer and thus to obtain a tentative relationship between this coefficient and vertical variation of σ_t. The results are based on data given in table 8 for stations 1172 (11°43′N, 40°54′W), 1174 (8°20′N, 40°45′W), and 1176 (5°16′N, 39°47′W). At these stations the temperature discontinuity is very highly developed between depths of 100 to 180, 30 to 100, and 110 to 200 meters, re-

[29] The calculation was based on the theoretical annual consumption of 52 cc oxygen per square centimeter of surface in this region, and on a carbon-phosphorus ratio of $1:2.73 \times 10^{-2}$ for oceanic plankton (Seiwell, 1934, 1935).

[30] Because of the general low phosphate content of the 0–50 meter layer in the area of investigation (page 35) it does not appear that a direct supply of phosphate by horizontal currents could be significant.

[31] Corrected for salt error; see page 7.

spectively (compare with table 7 for position of transport layer at these stations). The average vertical variation of σ_t per centimeter in this part of the discontinuity layer (2.81×10^{-4} to 3.00×10^{-4}) is approximately equal to a stability of $E = 3000 \times 10^{-8}$; and the average vertical variation of phosphate, 4.22×10^{-12} to 5.43×10^{-12} g/cm³/cm (average $= 4.88 \times 10^{-12}$), is 5 to 15 times as great as the average for the whole of the transport layer throughout the area.

TABLE 8

Station	Depth Range of Discontinuity	P Range[32] Mg/M³	σ_t Range	$\frac{\Delta\sigma_t}{\Delta Z}$	$\frac{\Delta P}{\Delta Z}\ \frac{\text{g/cm}^3}{\text{cm}}$
1172	100–180	1–41	24.35–26.60	2.81×10^{-4}	5.00×10^{-12}
1174	30–100	7–45	24.38–26.35	2.81×10^{-4}	5.43×10^{-12}
1176	110–200	5–43	24.07–26.77	3.00×10^{-4}	4.22×10^{-12}

Data for calculation of relationship of eddy conductivity and stability (see text, page 54).

The amount of phosphate (N) transported through the strongly marked discontinuity layer by an average eddy conductivity coefficient of 2 C.G.S. units is:

$$N = A\frac{\Delta P}{\Delta Z} = 2 \times 4.88 \times 10^{-12} = 9.76 \times 10^{-12}\ \text{g/cm}^2\text{/sec}$$

$$9.76 \times 10^{-12} \times 31.5 \times 10^{6} = 30.7 \times 10^{-5}\ \text{g/cm}^2\text{/year}$$

which is about 40 per cent of the total amount of phosphate calculated to be consumed annually in the overlying surface layer (page 45). Also, since this amount of phosphate is assumed to be the net eddy transport through the whole transport layer (at identical positions), the average eddy coefficient calculated for this layer is:

$$A = \frac{N}{\Delta P/\Delta Z} = \frac{9.76 \times 10^{-12}}{10.7 \times 10^{-13}} = 9.1\ \text{C.G.S. units}$$

where N is the amount of phosphate transported per unit of surface area (grams per square centimeter of horizontal surface per second) and $\Delta P/\Delta Z$ the average vertical variation of phosphate in the transport layer of stations 1172, 1174, and 1176 (table 7). The eddy coefficient, 9.1 C.G.S. units, is assumed to be characteristic of a vertical variation of 5.20×10^{-5} σ_t units per centimeter, which is the average σ_t variation for the transport layer of stations 1172, 1174, and 1176 (table 7). Hence, on the hypothesis that within the region of investigation (except as noted on page 43) the eddy coefficient is inversely proportional to the stability, we have:[33]

$$A : 5.20 \times 10^{-5} :: 9.1 : \frac{\Delta\sigma_t}{\Delta Z}$$

$$A = \frac{4.73 \times 10^{-4}}{\Delta\sigma_t/\Delta Z}$$

[32] Corrected for salt error; see page 7.
[33] This relationship is allowable for the order of magnitude of the stabilities dealt with in this paper, but cannot be extended to very low stabilities as A will then approach infinity; see footnote 27.

where 5.20×10^{-5} is the average vertical variation of σ_t per centimeter in the transport layer for stations 1172, 1174, and 1176; and $\Delta\sigma_t/\Delta Z$ that in the transport layer of any other part of the region.

The values of the eddy coefficients in the transport layer for individual stations thus calculated are given in table 7 (page 44). Because of the nature of the fundamental data on which this calculation rests (page 45) results are subjected to an indefinite error so that by using average values based on data in table 7, individual variations at stations are minimized and a better picture of regional variation of eddy conductivity is obtained. Thus, Fig. 25 indicates that, if regional variation of the eddy conductivity coefficient is the reverse of the vertical variation of σ_t, the average eddy coefficient for the transport layer is calculated to be about 30 C.G.S. units north of latitude 30°N and 10 C.G.S. units or less south of latitude 15°N.

Calculation of Vertical Upward Transport of Phosphate from Rich Midstrata to Impoverished Surface Layer

Using the results from the preceding discussion we may now estimate the vertical transfer of phosphate, per unit of time, from the rich midstrata to the phosphate poor layer overlying the thermocline by the following equation:

$$N = A\frac{\Delta P}{\Delta Z} \times 31.5 \times 10^6$$

where N represents grams of phosphate transferred to the homogeneous layer per square centimeter of surface per year; A, the eddy coefficient estimated above (page 46), and $\Delta P/\Delta Z$ the average phosphate gradient in the transport layer (page 43). The results, averaged for 5° coordinate squares in figure 25, show that the magnitude of transfer in the whole area is confined to narrow limits, between 3×10^{-4} and 6×10^{-4} grams of P per square centimeter of horizontal surface area per year; the higher values (5×10^{-4} to 6×10^{-4}) occur north of latitude 30°N. Thus, the higher concentrations of phosphate (with correspondingly greater vertical variations with depth) found in the southern half of the area (figs. 6, 7) appear to be offset in their influence on eddy transport of phosphate by proportionately larger vertical variations of σ_t and consequently smaller eddy coefficients.

The significance of these results in relation to the fertility of the area depends on whether or not organic production is principally confined to the water overlying the thermocline. Thus, in the northern part of the area where the homogeneous phosphate poor layer is 300–500 meters thick (table 7) eddy transfer of phosphate from below is apparently an important factor for fertility, but in the southern part, where the discontinuity layer begins close to the surface, the possibility exists that phytoplankton may utilize directly phosphate from the rich layers as discussed on page 50.

We have so far been dealing with vertical transport of phosphate per unit of surface or horizontal area, but, as it is customary to measure substances in the sea in terms of either their weight per unit weight or weight per unit volume of water (see Kalle, 1935), e.g., mg P per m³, the foregoing results are likewise recalculated (figure 25). Thus, provided that, N, the amount of phosphate (grams) transferred per square centimeter of horizontal area be equally distributed throughout the entire thickness, ΔZ, of the homogeneous layer then:

$$N_v = \frac{N \times 10^7}{\Delta Z(\text{meters})}$$

where N_v is milligrams of phosphate supplied annually by eddy transfer per cubic meter of water overlying the transport layer. Results of this calculation for 5° coordinate squares, based on data in table 7, given in figure 25, show an annual gain of phosphate per cubic meter ranging from less than 15 mg P in the north to more than 60 mg P in the south; the greater gains per unit volume occurring where the discontinuity layer lies close to the surface.

According to these calculations, while eddy transfer of phosphate to the water overlying the thermocline is about one-half as much in the southern half of the area as in the northern half, the apparent enrichment of this layer per unit volume (milligrams of P per cubic meter) in the south may be more than 4 times that in the north. Hence, it appears that regional variations of chemical fertility[34] in the area cannot be estimated simply by comparing unit volume measurements of nutrient salt content at similar levels. Further, it seems likely that this conclusion applies for other parts of the oceans as well. Comparing the fertility of several regions based solely on the concentrations of products per unit volume of water at similar levels may lead to erroneous interpretation of conditions, particularly if the regions differ in vertical distribution of the physical and chemical qualities of the water.

On the basis of this discussion it may be tentatively concluded that, throughout the western North Atlantic (regardless of existing density gradients), there is a significant eddy transfer of phosphate from the rich midstrata to the impoverished surface layer. This is somewhat in opposition to previous ideas that a well developed discontinuity layer restricts vertical transfer of nutrient substances to such an extent that, for instance, the relatively great quantities of phosphate in the midstrata of the tropics are lost to the cycle of life for an indefinitely long time (Harvey, 1928; Wattenberg, 1927; Seiwell, 1931). On the contrary the results of the present analyses, suggest a mechanism by which exchange of nutrient substances may be brought about in very stable waters, and indicate that in the ocean basins eddy transfer may be a significant factor in determining fertility.

Since, within the area of investigation, phosphate supply to the surface layer appears to depend principally both on transfer from below and on regeneration of phosphate within the layer itself, it follows that the importance of eddy transfer in fertility of a given region depends on the extent to which phytoplanktonic activity is confined to the phosphate poor water overlying the discontinuity layer. This, at present, is an open question because the north south rise of the discontinuity layer toward the surface (fig. 6) may bring phosphate rich water within direct reach of phytoplankton, unless the depth of phytoplanktonic activity is limited by some factor other than light.

[34] The term chemical fertility is used here to indicate the capability of a water column of unit area in the photosynthetic layer to produce organic matter on the basis of its nutrient salt content. In general, much of this discussion of the circulation of phosphate can equally well be applied to that of any nutrient salt. And, as the question of a limiting nutrient substance for organic production does not enter in this discussion, the term chemical fertility may be used here as synonymous with phosphate fertility.

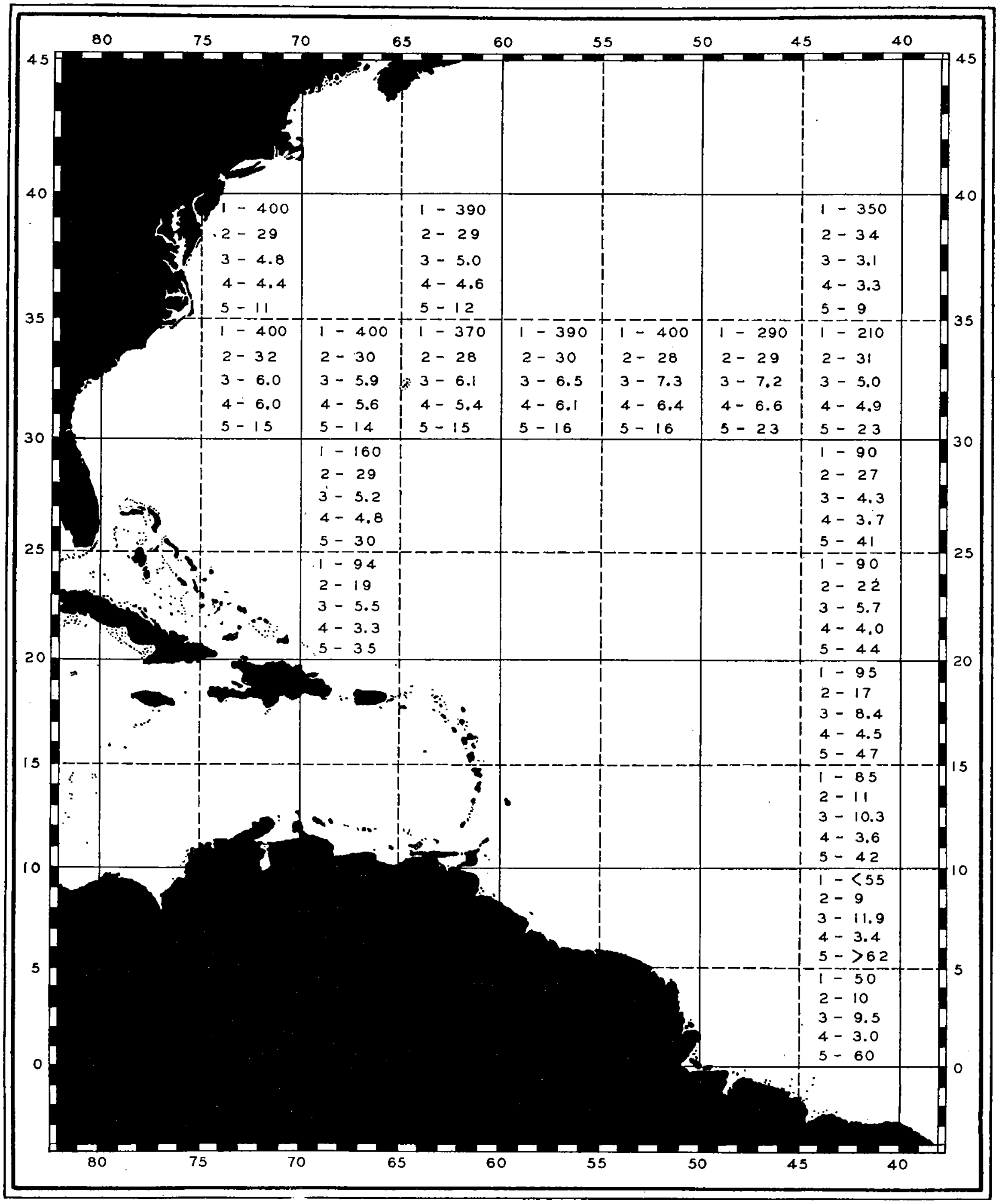

FIG. 25. Average values for five degree coordinate blocks calculated from data in table 7.
Column 1. Thickness of water layer overlying thermocline (page 42).
Column 2. Eddy conductivity coefficient for transport layer (page 45).
Column 3. Vertical variation of phosphate in transport layer (grams per cubic meter per centimeter$\times 10^{13}$; page 43).
Column 4. Amount of phosphate transported annually from transport layer to overlying water (grams of P per square centimeter of surface$\times 10^{4}$; page 47).
Column 5. Annual enrichment of water overlying transport layer by eddy conductivity (milligrams of P per cubic meter; page 48).

Depth of Layer of Plant Activity

Within the area of investigation the question of restriction of the depth of plant activity by the thermocline (page 9) increases in importance from north to south; for, if a significant consumption of phosphate should occur in the phosphate rich water of the thermocline of tropical latitudes the chemical fertility would be undoubtedly independent of phosphate transferred upward by eddy conductivity. Indeed, it seems that if light alone determines the depth to which plant growth in low latitudes extends (as generally supposed) there would be available for organic production in certain parts of

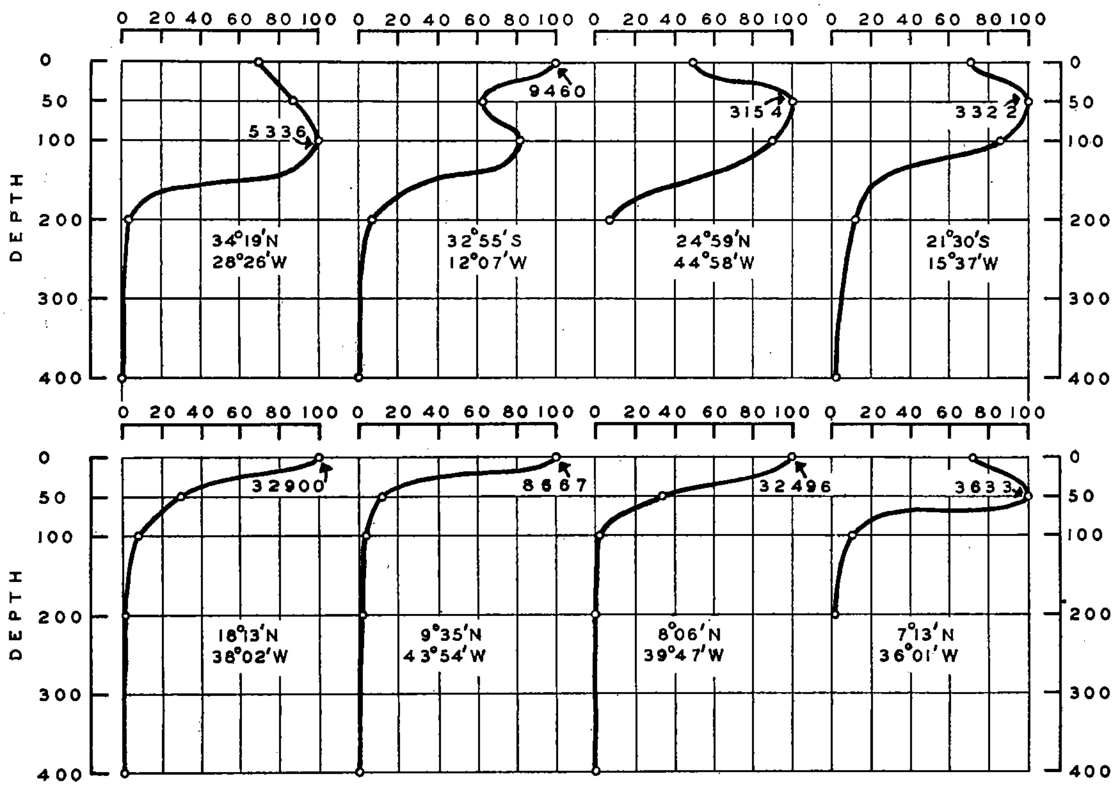

Fig. 26. Relative percental vertical distribution of unicellular plants in the Atlantic ocean. The observed maximum populations (cells per liter) used as a basis for calculation is given for each curve.

the tropical North Atlantic basin a supply of nutrient substances (if we can judge from the vertical distribution of phosphate; figs. 6, 7) as great even as occurs in the surface layers at high latitudes in the South Atlantic (Deacon, 1933; Ruud, 1930) or off the northwest coast of Africa (Wattenberg, 1927).

On the other hand, the vertical distribution of phosphate and oxygen in the upper part of the water column of the area investigated suggests in general that the depth of plant activity is limited by the thermocline, even in low latitudes where the latter extends well up into the illuminated part of the water column. And this idea, is, at least, not refuted by the few results of phytoplanktonic investigations in the Atlantic.

Unfortunately, at the present time, there is no evidence to quantitatively illustrate regional depth variations of phytoplanktonic activity in the North Atlantic ocean. Because of the difficulties encountered in obtaining a true sample of the entire phytoplanktonic population, and because organisms taken from subsurface strata may be in an inactive state due to sinking from higher levels such data must, perhaps, be obtained from culture experiments directly on the high seas (used and described by Gran, 1927, 1933). Figure 26, illustrating regional variation in vertical distribution of unicellular plants for

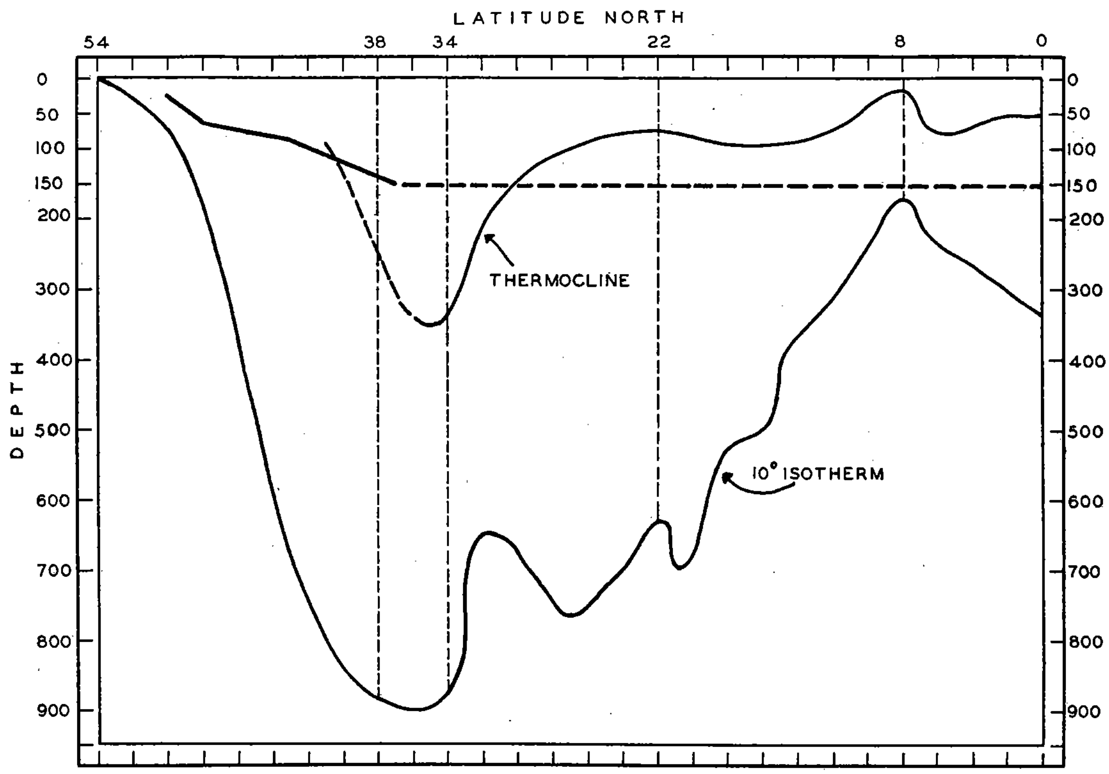

FIG. 27. Diagrammatic sketch showing estimated depth of photosynthetic layer, depth of 10° isotherm and thickness of water overlying thermocline for mid Atlantic between 50°N and equator. Depth of photosynthetic layer based on Clark's (1933) measurements of depth at which irridation of surface light was reduced to one per cent. Between 37°N and equator there are no light penetration measurements (so far as I know) so it is assumed since the Sargasso Sea is the most transparent part of the ocean investigated by Oster and Clarke (1935) that there is little difference between it and the equator.

various positions in the North and South Atlantic, has been prepared with these considerations in mind. The curves are based on phytoplanktonic population of centrifuge samples published by Lohman (1920) and Hentschel (1932) for the North and South Atlantic; the population for each depth has been recalculated in terms of percentage of the maximum observed population in the vertical distribution; the greatest number of cells per liter (assumed to be 100 per cent) is entered on each curve at the depth where it occurs.

So far as they go these curves indicate, in a general way, that in the lowest latitudes of the North Atlantic the bulk of the phytoplanktonic population is confined to a more narrow layer than it is further north so that there is at least a general relationship between the depth of phytoplanktonic activity and the surfaceward approach of the discontinuity layer. This together with the evidence above appears to support the idea that temperature instead of light limits the depth of plant activity in the tropical North Atlantic.

Temperature may bar access, to tropical latitudes, for plant species adjusted to low temperatures (not higher than 10°–12°), because the great thickness of the overlying warm water in mid latitudes (figs. 8, 27) would (for lack of light) prevent a continuous planktonic distribution between surface strata in high latitudes and the underlying strata of similar temperature in the tropics. Phytoplanktonic communities in the tropics would thus tend to be limited to "high temperature" species; and the lower and cooler parts of the tropical photic layer would be left barren of plants unless some development of "low temperature" organisms had taken place there independent of any direct connection with similar "low temperature" communities in high latitudes to the north or south.

Oceanic planktonic observations of Lohman (1920) and Hentschel (1932), both in North and South Atlantic, do in fact show a change in the phytoplanktonic community between mid and low latitudes, and it is perhaps significant that the preponderance of diatoms in high latitudes is to a large extent replaced by blue green algae in the tropics. Lack of knowledge of the light requirements of marine plants and of light penetration into the sea prevents more definite conclusions. Indeed it is also possible that light alone may be effective in reducing the thickness of the photosynthetic layer of the tropics, such a condition as might be caused by an intense development of phytoplankton close to the surface forming an effective screen to the deeper penetration of light (see Moberg, 1928). However, in the absence of light penetration measurements from the tropical Atlantic reduction in thickness of the photosynthetic layer by temperature seems the more probable.

BIBLIOGRAPHY

Atkins, W. R. G.

1923 The phosphate content of fresh and salt waters in its relationship to the growth of the algal plankton. Journ. Mar. Biol. Assoc., Vol. XIII. pp. 119–150.

1926(a) A quantitative consideration of some factors concerned in plant growth in water. Part II. Some chemical factors. Journ. du Conseil., Vol. I, No. 3. pp. 29–58.

1926(b) The phosphate content of sea water in relation to the growth of the algal plankton. Part III. Journ. Mar. Biol. Assoc., Vol. XIV. pp. 447–467.

Bigelow, Henry B.

1933 Studies of the waters on the continental shelf, Cape Cod to Chesapeake Bay. I, The cycle of temperature. Papers in Phys. Ocean. and Meteorol., Vol. II, No. 4. 135 pp.

Böhnecke, Günther; Føyn, Birgithe; und Wattenberg, Hermann

1932 Beiträge zur Ozeanographie des Oberflächenwassers in der Dänemarkstrasse und Irminger See. II Teil. Annal. der Hydrog. und Marit. Meteorol., Vol. LX. pp. 314–321.

Braarud, Trygve and Klem, Alf.

1931 Hydrographical and chemical investigations in the coastal waters off Møre and in the Romsdalsfjord. Skrifter utgitt av Hvalraadet ved Univ. Biol. Laborat. Oslo, No. 1. 88 pp.

Brandt, K.

1899 Ueber den Stoffwechsel im Meere. Wiss. Meeresuntersuch. abt., Kiel, Vol. 4, pp. 215–230.

1920 Über den Stoffwechsel im Meere. 3 abhandlung. Wiss. Meeresuntersuch. abt. Kiel, Vol. 18. pp. 185–430.

Brujewicz, S. W. and Krasnova, V. S.

1933 Salt and temperature corrections for the estimation of phosphates in sea water. Bull. 14 of State Ocean. Instit. of U.S.S.R., pp. 33–43.

Buch, Kurt.

1932 Untersuchungen über gelöste phosphate und stickstoffverbindungen in den Nordatlantischen meeresgebieten. Merentutkimuslaitoksen Julkaisu Havsforskningsinstitutets Skrift, N:o 86. 30 pp.

Clarke, George L.

1933 Observations on the penetration of daylight into mid-Atlantic and coastal waters. Biol. Bull. Vol. LXV, No. 2. pp. 317–337.

Conseil Permanent International pour L'Exploration de la Mer.

1933 Bull. Hydrog. pour L'Année, 1932.

Deacon, G. E. R.

1933 A general account of the hydrology of the South Atlantic ocean. Discovery reports, Vol. VII. pp. 171–238.

Dehn, William M.

1917 Fallacies in colorimetry. Journ. Amer. Chem. Soc., Vol. XXXIX, No. 7. pp. 1392–1398.

Discovery Report.
1932 Station list, 1929–1931. Vol. IV. pp. 1–232.
Gran, H. H.
1927 The production of plankton in the coastal waters off Bergen March–April 1922. Rept. Norwegian Fish. and Marine Invest., Vol. III, No. 8. 74 pp.
1932 Phytoplankton, methods and problems. Journ. du Conseil, Vol. VII, No. 3. p. 343–358.
1933 Studies on the biology and chemistry of the Gulf of Maine. II. Distribution of phytoplankton in August, 1932. Biol. Bull., Vol. LXIV, No. 2. pp. 159–182.
Gran, H. H. and Thompson, Thomas G.
1930 The diatoms and the physical and chemical conditions of the sea water of the San Juan Archipelago. Publ. Puget Sound Biol. Stat., Vol. 7. pp. 169–204.
Harvey, H. W.
1928 Biological chemistry and physics of sea water. University Press, Cambridge. 194 pp.
1929 Methods of estimating phosphates and nitrates in sea water. Rapp. et Proc. Verb., Vol. LIII. pp. 68–74.
Helland-Hansen, Bjørn.
1930 Physical oceanography and meteorology. Report of the Scientific Results of the "Michael Sars" North Atlant. Deep Sea Exped. 1910, Vol. I, Part I (text). 115 pp. Part II (tables and plates). 102 pp.
Hentschel, Ernst.
1932 Die Biologischen Methoden und das Biologische Beobachtungsmaterial der "Meteor" Expedition. Wiss. Ergeb. Deutschen Atlant. Exped. "Meteor" 1925–1927, Vol. X. 274 pp.
Hesselberg, Th. und Sverdrup, H. U.
1915 Die Stabilitätsverhältnisse des Seewassers bei vertikalen Verschiebungen. Bergens Museums Aarbok 1914–1915, No. 15, 16 pp.
Ibañez, O. G.
1933 Note on the effect of salts in the determination of phosphate in sea water by Denigès method. Journ. du Conseil., Vol. VIII. pp. 326–329.
Kalle, Kurt.
1934 Meereskundliche chemische Untersuchungen mit Hilfe des Zeissschen Pulfrich-photometers. III Mitteilung. Methodische Untersuchung der Phosphatgehaltsbestimmung. Ann. der Hydro. u. Marit. Meteorol., Vol. LXII. pp. 95–102.
1935 Zur Frage der Festsetzung der Masseinheiten in der meereskundlichen Chemie. Ann. der Hydro. u. Marit. Meteorol., Vol. LXIII, No. 3. pp. 18–23.
Knudsen, Martin.
1901 Hydrographical tables. Copenhagen. 63 pp.
Lohmann, H.
1919 Die Bevölkerung des Ozeans mit plankton nach den Ergebnissen der Zentrifugenfänge während der Ausreise der "Deutschland" 1911. Archiv für Biontologie, Vol. IV, No. 3, Part 1. pp. 1–470. Part 2. pp. 471–617.
Matthews, D. J.
1917 On the amount of phosphoric acid in the sea water off Plymouth Sound. Journ. Mar. Biol. Assoc., Vol. 11. pp. 251–257.

McEwen, George F.
1919 Ocean temperatures, their relation to solar radiation and oceanic circulation. Misc. studies in agric. and biol. In Semicentennial Publ. Univ. Calif. 1868–1918. pp. 335–421.
Moberg, Erik G.
1928 The interrelation between diatoms, their chemical environment, and upwelling water in the sea, off the coast of southern California. Proc. Nat. Acad. of Sci., Vol. 14, No. 7. pp. 511–518.
Moberg, Erik G.; Seiwell, H. R.; Graham, H. W.; Paul, J. H.
1930 The phosphate content of the surface water in the Pacific as related to the circulation. Reports and Communications, Sect. Terr. Mag. Elec., Sec. Oceanography, Intern. Geod. Geophys. Union, Washington, D.C. 1930. pp. 98–100.
Oster, R. H. and Clarke, G. L.
The pernetration of the red, green, and violet components of daylight into Atlantic waters. (Data obtained from galley proof; to be published in Journ. Opt. Soc., 1935.)
Raben, E.
1920 Quantitative Bestimmung der im Meerwassergelösten phosphorsäure. Wiss. Meeresuntersuch. Abt. Kiel, Vol. 18. pp. 1–24.
Ruud, Johan, T.
1930 Nitrates and phosphates in the southern seas. Journ. du Conseil., Vol. V, No. 3. pp. 347–360.
Schmidt, Dr. Wilhelm.
1925 Der Massenaustausch in freier Luft und verwandte Erscheinungen. Probleme der Kosmischen Physik. VII. Hamburg. 118 pp.
Seiwell, H. R.
1931 Observations on the phosphate content and hydrogen ion concentration of the North Sea, the southern entrance to the Norwegian Sea, and the water south of Iceland. Journ. du Conseil., Vol. VI, No. 2. pp. 213–231.
1931 A consideration of some external factors governing the production of plankton in the sea. Journ. of Ecology, Vol. XIX, No. 1. pp. 164–176.
1934 The distribution of oxygen in the western basin of the North Atlantic. Papers in Phys. Ocean. and Meteorol., Vol. III, No. 1. 86 pp.
1935 The annual organic production and nutrient phosphorus requirement in the tropical western North Atlantic. To be published in Journ. du Conseil., Vol. X, No. 1.
Seiwell, H. R. und Seiwell, G. E.
1934 Über den Gesamtphosphorgehalt des Seewassers im westlichen Nordatlantischen Ozean. Annal. der Hydro. und Marit. Meteorol., Vol. LXII, pp. 7–13.
Sverdrup, H. U.
1933 Scientific results of the "Nautilus" expedition, 1931. II. Oceanography. Papers in Phys. Ocean. and Meteorol., Vol. II, No. 1. pp. 16–63.
Thomsen, Helge.
1931 Nitrate and phosphate contents of Mediterranean water. Rept. on Danish Oceanographical Expeds. 1908–10 to Medit. and adjacent seas, Vol. III, No. 6. 14 pp.

Wattenberg, H.
1927 Bericht über die chemischen Arbeiten. Die Deutsche Atlant. Exped. "Meteor", III Bericht. pp. 137–143.
1933 Das chemische Beobachtungsmaterial und seine Gewinnung. Wiss. Ergeb. Deutschen Atlant. Exped. "Meteor" 1925–1927, Vol. VIII. pp. 9–121.